

4000 years of Environmental and Climate Change at Eilandvlei:
A palynological investigation into the Late Holocene palaeoenvironment of the
Wilderness Embayment

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Abstract

It is a well-known fact that the South African palaeoenvironmental record is fairly limited, mainly because the environmental conditions here do not lend themselves to the preservation of proxy data sources. This being said, new evidence is emerging from the Wilderness Embayment along the Southern Cape coast of South Africa. This area is of particular interest from a palaeoclimate perspective due to its location within the small year round rainfall zone of South Africa, while the presence of both Fynbos and Afrotemperate forest within this region further highlights the importance of studies in the area. This study presents a new high resolution late Holocene pollen and charcoal record from Eilandvlei, extending from c. 3 800 cal yr BP to present. The record is marked by several significant periods – largely corresponding to the timing of the three events that typify the late Holocene: the Neoglacial, the Medieval Climate Anomaly and the Little Ice Age.

The record also encompasses the arrival of European colonists in the region ca. 1800 AD, marked by the appearance of *Pinus*, with subsequent notable fluctuations in other taxa, e.g. *Podocarpus* and *Stoebe*-type, possibly the expression of the anthropogenic effect on the landscape. This high resolution record thus highlights significant, albeit short term, fluctuations in climate and vegetation patterns along the Southern coast of South Africa and adds to the inadequate records available for the region thus far.

Acknowledgements

Simply putting words on paper will never be able to convey my gratitude to those who have supported and carried me throughout this process. I never thought this day would finally arrive.

To my supervisors, Mike and Lynne, and my family, mom, Mia and John – thank you

This thesis is dedicated to my father

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1. Introduction

1.1. Why, how, when and where?

In this era where there is increased focus on future environmental and climate change, an essential question comes to the forefront: How can we use the past to predict the future? Indeed, Gil-Romera et al. (2010, page 43) state that “The need to understand past environmental fluctuations to improve our knowledge on current global change has been widely acknowledged”. In this regard, Meadows (2001, 2012) highlights the late Quaternary in particular as a significant period against which contemporary, and future, environmental and climate change can be measured. This idea is shared by MacDonald et al. (2008) stating that Quaternary climate change provides the best analogue for understanding contemporary climate change and the possible, or expected, future impacts.

Since the 1920's the most popular, and widely applied, method for the investigation of late Quaternary environments is pollen analysis, or palynology (Faegri & Iversen 1989). This discipline involves the study of pollen grains and spores and focuses on the structure and formation of pollen and spores, as well as their preservation and dispersal (Moore et al. 1991). The focus of Quaternary palynology then, is the palaeoecological analysis of fossil pollen and spores (MacDonald 1988) preserved in a range of environments, over time. These preserved fossils can offer information on vegetation, and the associated ecosystem changes, at certain time periods in the past. This technique is also applicable in the study of past climate change, although certain concerns have to be taken into account; these include the ‘human impact’, the lagged response of vegetation to changes in climate and the possible influence of the local soil types (Moore et al. 1991).

Palynology has been extensively used in the study, and reconstruction, of vegetation histories with a significant focus on the Holocene (Moore et al. 1991) – the time period spanning the last c. 11 500 cal years. The Holocene is characterised by distinctive and globally contemporaneous climatic events, including the Holocene Altithermal, the Neoglacial, the Medieval Climate Anomaly and the Little Ice Age. The latter three have been termed “rapid climate change” events by Mayewski et al. (2004) and typifies the Late Holocene – the last c. 4200 cal years (Walker et al. 2012).

The South African palaeoenvironmental record is fairly limited, mainly due to the fact that environmental conditions here do not lend themselves to the preservation of proxy data sources (Chase & Meadows 2007). This being said, there is one region, in particular, in South Africa which provides a more detailed record of Quaternary change – the Southern Cape. This region, especially

along the coast, has been the subject of a variety of archaeological studies, including Pinnacle Point, near Mossel Bay (e.g. Brown et al., 2009; Jerardino and Marean, 2010; Marean, 2010; Brown et al., 2012), and Nelson Bay Cave, in the vicinity of Plettenberg Bay (e.g. Inskeep, 1972; Klein, 1972; Fairhall and Young, 1973). In relation to the archaeological records, palaeoenvironmental records are fairly limited, with the most prominent including those from the Wilderness Embayment by Martin (1956, 1959, 1960, 1968), and to a lesser extent, the palynological record from the Norga peat, near George (Scholtz 1986). This is changing however, with new records from the Wilderness Embayment recently been made available (e.g. Kirsten 2014; Reinwarth et al. 2013).

Situated within the year round fall zone, studies in the Southern Cape can provide great insight into the changing dynamics of both the winter- and summer rainfall zones during the late Quaternary. Furthermore, the Wilderness Lakes provide the perfect environment for the accumulation and preservation of palaeoenvironmental proxy material, including pollen. Consequently, this region represents a rich archive of (largely untapped) palaeoenvironmental evidence and holds great potential for future research.

1.2. Research Questions

This study explores Holocene climate and vegetation change in the Wilderness Embayment, using palaeoenvironmental evidence from one of the lakes in this embayment, Eilandvlei, with the purpose of answering the following:

- How has the environment of the Wilderness Embayment changed during the late Holocene in terms of vegetation?
- How has the climate in the year round rainfall changed during the late Holocene and what are the driving mechanisms behind these changes?
- What is the timing of these changes in climate and how does it relate to changes in vegetation at Eilandvlei, and thus the Wilderness Embayment?
- How does the late Holocene environmental history of the Wilderness Embayment, as inferred at Eilandvlei, compare to local and regional palaeoenvironmental records, as well as with records from the winter- and summer rainfall zones?

1.3. Aims and Objectives

The central aim of this project is to reconstruct a late Holocene record of environmental and climate change for Eilandvlei and its associated catchment, by means of pollen- and microscopic charcoal analysis.

A secondary aim is to establish the relative importance of various environmental drivers that may have affected the late Holocene environment in this area. More specifically, the drivers to be investigated are: climate change, sea level fluctuations and the human- or anthropogenic impact.

In order to reach these aims, the following objectives are put forward:

1. The extraction of sediment cores from the specified study site, i.e. Eilandvlei
2. Sub-sampling of these cores to obtain:
 - High resolution chronologies through radiocarbon dating
 - The palynological record of vegetation change by means of pollen analysis, which includes the extraction, counting and identification of pollen grains in the samples
 - The charcoal record, associated with the pollen record, to examine fire regimes in the area
3. Examination of these results in order to reconstruct palaeoenvironmental conditions in the area
4. From this, investigate the importance of the aforementioned environmental drivers
5. Evaluate these findings in the context of palaeoenvironmental records from South Africa

1.4. Thesis Outline

The following chapter introduces the Southern Cape, describing the contemporary regional climate and vegetation, as well as its geological and geomorphological characteristics. This chapter further presents the study area, Eilandvlei and the Wilderness Embayment.

In Chapter 3, a review of selected records representing the three rainfall zones of South Africa are provided.

Chapter 4 outlines the methods employed in obtaining palaeoenvironmental proxy evidence - from fieldwork to the statistical analysis of the data.

The results for Eilandvlei are presented in Chapter 5. These are: the chronological results, the stratigraphic description of the core as well as the results from the pollen- and micro-charcoal analyses.

Chapter 6 discusses these results in terms of the climatic and environmental significance of major indicator taxa to provide a picture of late Holocene changes in the Wilderness region. Significant events identified in the record are also further explored in this chapter, followed by the assessment of the Eilandvlei record in relation to the palaeoenvironmental records introduced in Chapter 3.

Lastly, Chapter 7 offers a summary of the findings from the Eilandvlei record and a review of the research aims and objectives.

2. The Southern Cape

2.1. Introduction

In exploring the contemporary environment of the Southern Cape, an understanding of both short- and long term changes, and the dynamics contained therein, in the region can be reached. This chapter thus presents an overview of the contemporary regional climate and vegetation, as well as the underlying geology of the Southern Cape.

2.2. Regional Setting

For the purpose of this study the Southern Cape is geographically defined as the area between Cape Agulhas and Cape Recife along the coast, with the Cape Fold Belt Mountains forming the northern boundary.

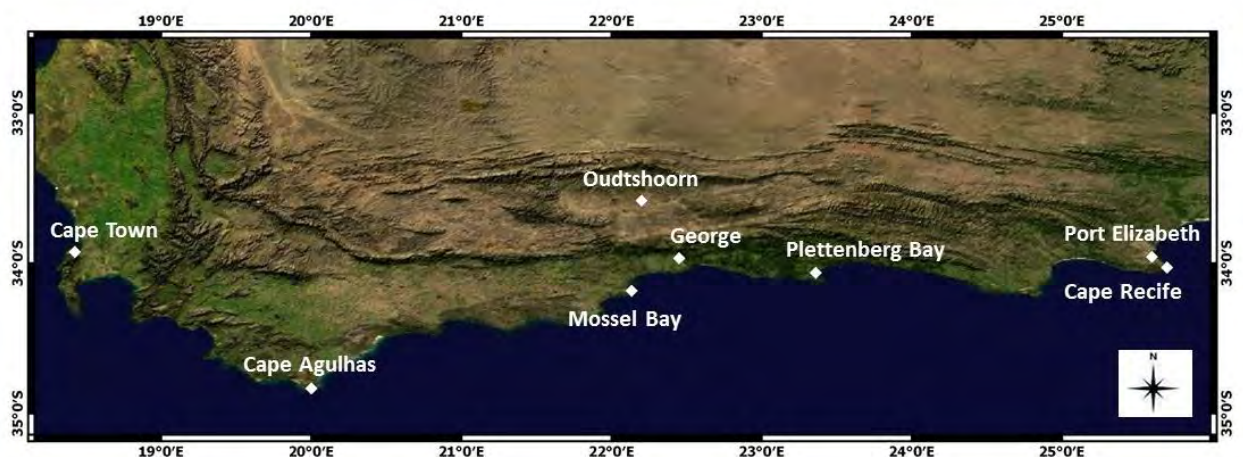


Figure 2.1. The Southern Cape (Cape Town is included as reference point)

2.3. Climate

2.3.1. Rainfall Seasonality

South Africa can broadly be divided into three rainfall zones. The winter rainfall zone (WRZ) is focussed on the West and South western coast while the summer rainfall zone (SRZ) incorporates the largest part of the interior and the East Coast. At the convergence of these, the smaller intermediate year round rainfall zone (YRZ) is found, centred on the Southern Cape coast.

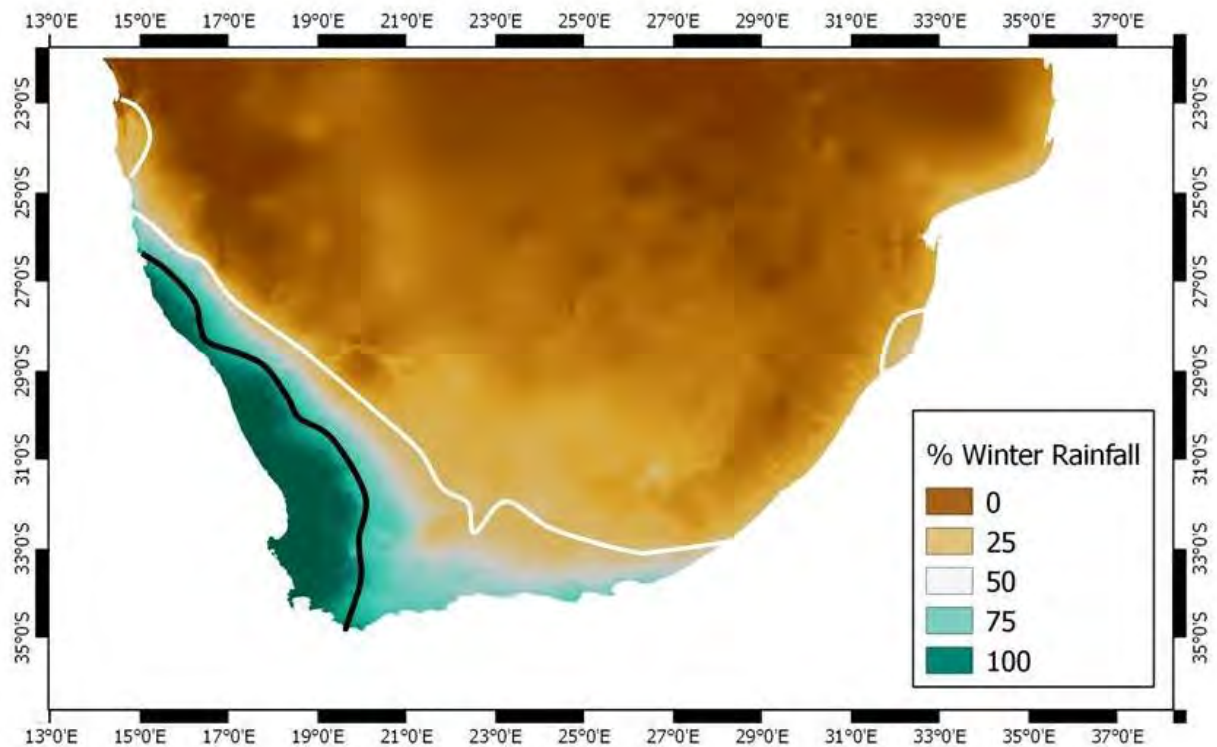


Figure 2.2. Rainfall seasonality across South Africa: the winter rainfall zone (WRZ) receives > 66% winter rain, the year round rainfall zone (YRZ) receives between 66 and 33% winter rain, and the summer rainfall zone (SRZ) receives < 33% winter rain (*sensu* Chase & Meadows 2007)

WRZ climate is influenced by the circumpolar westerlies located southward of the subcontinent (Tyson 1997; Tyson & Preston-Whyte 2000). During the austral winter, June to August, these westerlies intensify and expand. With this expansion, temperate frontal systems, embedded in the westerlies, transport moisture to the subcontinent resulting in rainfall over the region (Tyson & Preston-Whyte 2000).

In the SRZ, climate is principally affected by the interactions of easterly flows, connected to the Intertropical Convergence Zone (ITCZ), and subtropical high-pressure cells (Tyson & Preston-Whyte 2000). Greater easterly flow is observed from October to March as a result of the southward movement of the global circulation systems (Tyson & Preston-Whyte 2000). A semi-stationary low-pressure system further develops over the continent with moisture from the Indian Ocean, transported by the easterlies, resulting in rainfall in the SRZ (Tyson & Preston-Whyte 2000).

As the name implies, the YRZ receives rainfall during both summer and winter - largely affected by the winter frontal depressions embedded in the westerlies as well as monsoon rainfall and easterly winds associated with the SRZ (Cohen & Tyson 1995; Tyson & Preston-Whyte 2000).

2.3.2. Regional Southern Cape Climate

Formally, using the Köppen (1931) climate classification system, the coastal area between Cape Agulhas and George can be classified as BSk – semi-arid cool and dry, with the area from George to Port Elizabeth categorised as Cfb – wet all seasons, summers long and cool (Schulze et al. 2007).

In addition to its setting within the YRZ, several regional factors exert a further significant influence on the climate of the Southern Cape Coast:

Topography

The presence of the Cape Fold Belt and the Outeniqua Mountains have an important effect on climate and rainfall in the region. The Cape Fold Belt prevents the advance of moisture towards the north causing a rain shadow effect behind these mountains (Martin 1968; Hunter 1987). The Outeniqua Mountains in turn, are responsible for orographic uplift and atmospheric flux causing light rain, foggy and cloudy conditions as well as high relative humidity along the coast (Martin 1968; Hunter 1987).

The Agulhas Current

Southern Cape climate is additionally influenced by its location near the convergence of the Atlantic and Indian Oceans as well as the meeting of the Benguela and Agulhas ocean currents (Cohen & Tyson 1995; Bateman et al. 2004). The Agulhas current, specifically, has a major influence as climatic conditions along the coast are particularly susceptible to changes in the upwelling system along the eastern Agulhas Bank (Cohen & Tyson 1995; Bateman et al. 2004). Moreover, warm water from the Agulhas current provides a source of moisture, generating increased humidity along the coast (Cohen & Tyson 1995; Bateman et al. 2004). Scott and Lee-Thorp (2004) additionally suggest that this moisture supply is one of the more significant climatic parameters influencing changes in temperature in this region.

Bergwinds

Bergwinds follow a predominantly north-westerly to north-easterly direction from the drier interior, over the mountains towards the coast (Geldenhuys 1994). Most common during winter, bergwinds are linked to the movement of low pressure cells along the coast from west to east (Geldenhuys 1994). These winds are hot and desiccating resulting in large scale variability in temperature and moisture availability in the region (von Maltitz et al. 2003). In addition, bergwinds play a key role with regard to fire regimes, and consequently the distribution of forests, in the region (Geldenhuys 1994).

Rainfall and Temperature Trends

Although the southern Cape coastal plain (SCCP) generally receives year round rainfall, increased rainfall is experienced during early summer and autumn with average annual rainfall ranging from 500mm in the west and along the coast, to 1200mm in the mountain regions (von Maltitz et al. 2003).

Average maximum temperatures vary from 24 °C in summer, to 18 °C in winter, with average minimum temperatures ranging from 20 °C to 9 °C during the summer and winter, respectively (von Maltitz et al. 2003; Russel et al. 2012). Figures 2.3.1 to 2.3.7 illustrates rainfall and temperature trends in the region.

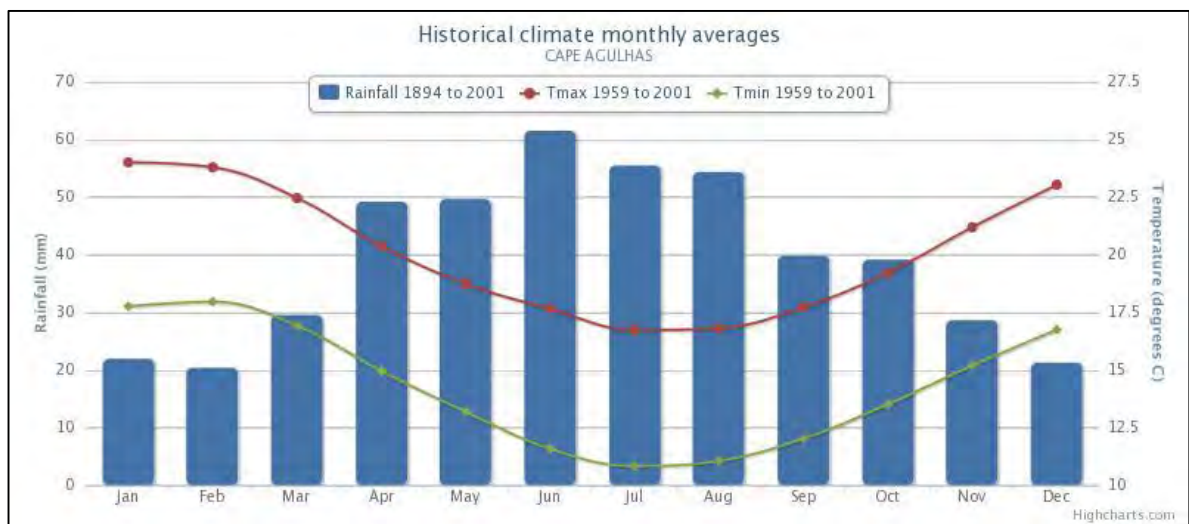


Figure 2.3.1. Trends in rainfall and temperature at Cape Agulhas (Source: CSAG, 2014; www.cip.csag.uct.ac.za)

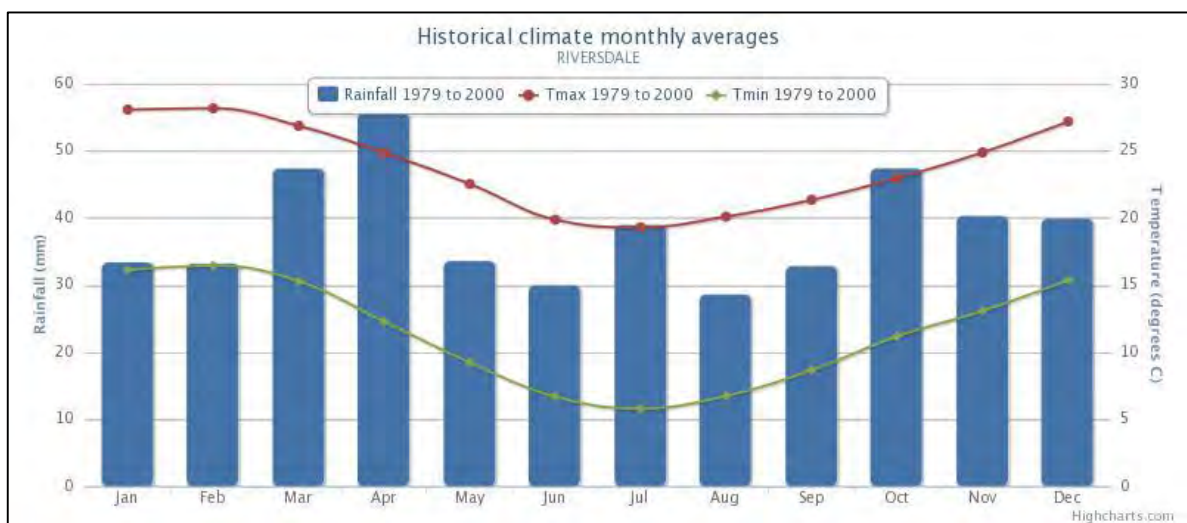


Figure 2.3.2. Trends in rainfall and temperature at Riversdale (Source: CSAG, 2014; www.cip.csag.uct.ac.za)

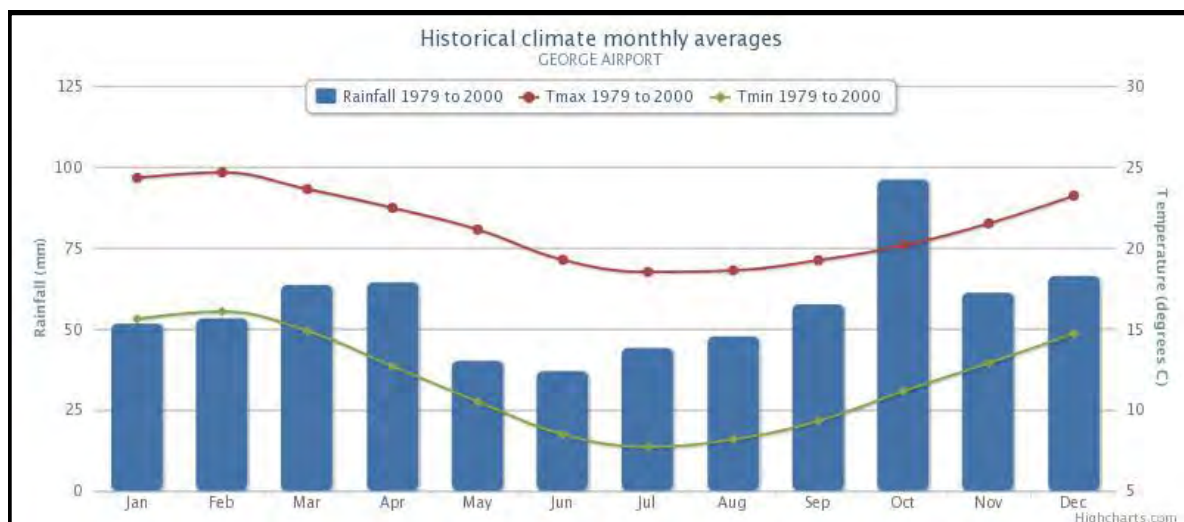


Figure 2.3.3. Trends in rainfall and temperature at George (Source: CSAG, 2014; www.cip.csag.uct.ac.za)

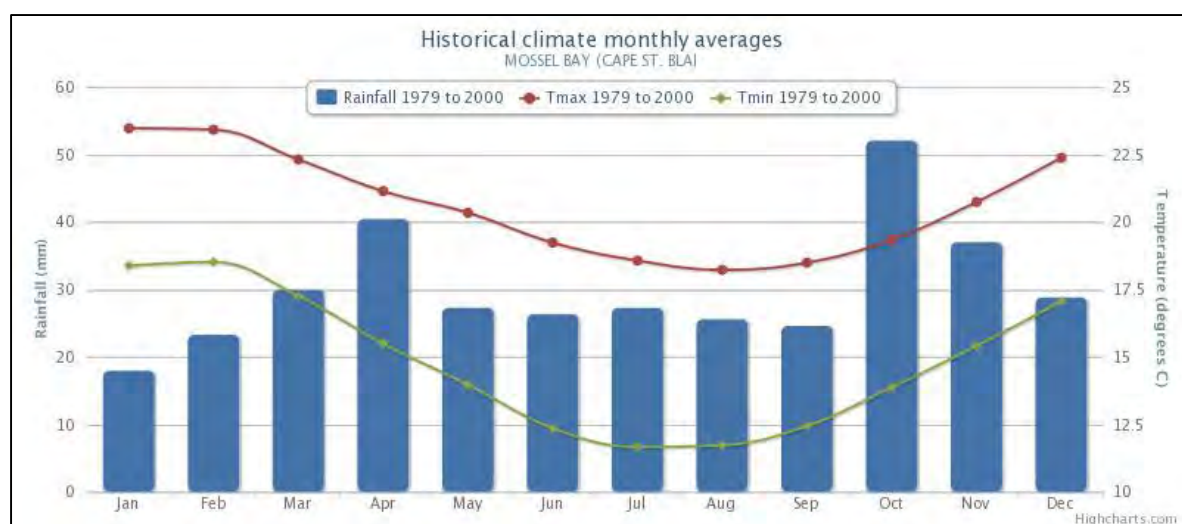


Figure 2.3.4. Trends in rainfall and temperature at Mossel Bay (Source: CSAG, 2014; www.cip.csag.uct.ac.za)

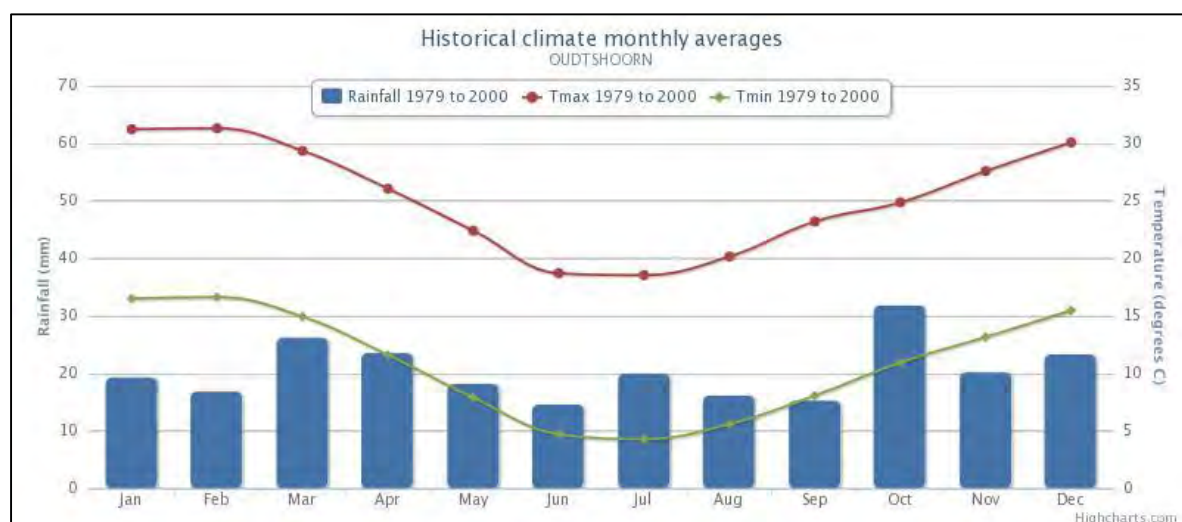


Figure 2.3.5. Trends in rainfall and temperature at Oudtshoorn (Source: CSAG, 2014; www.cip.csag.uct.ac.za)

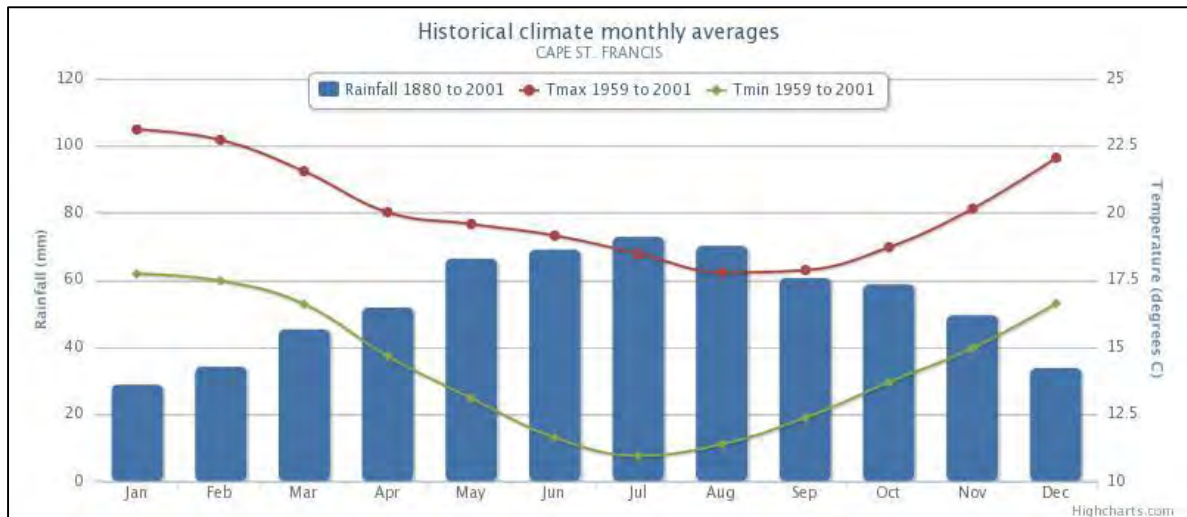


Figure 2.3.6. Trends in rainfall and temperature at Cape St. Francis (Source: CSAG, 2014; www.cip.csag.uct.ac.za)

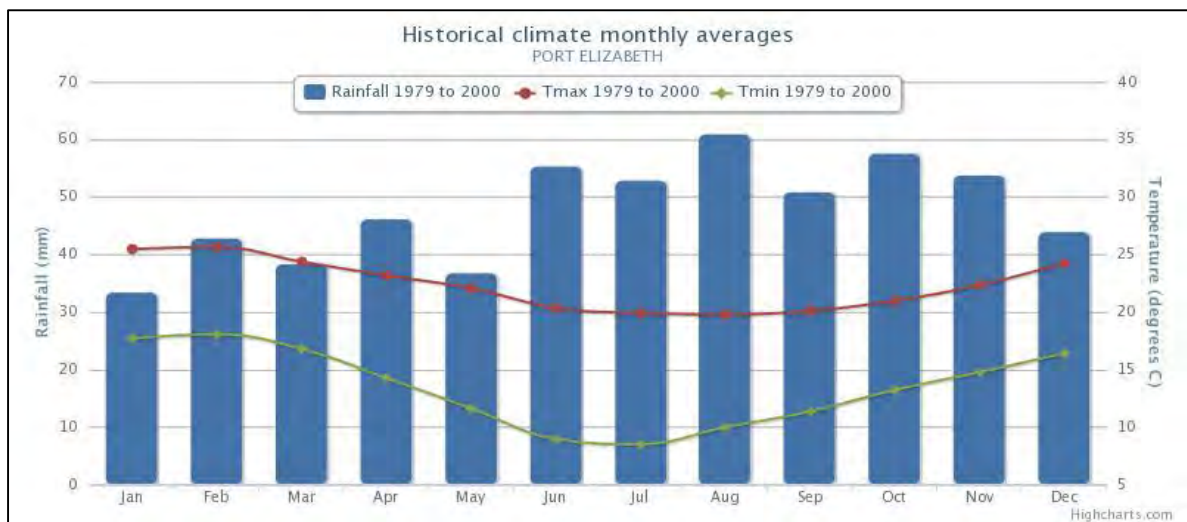


Figure 2.3.7. Trends in rainfall and temperature at Port Elizabeth (Source: CSAG, 2014; www.cip.csag.uct.ac.za)

2.4. Geology and Geomorphology

The Southern Cape coastal plain is thought to have originated from the break-up of Gondwana around 155 to 135 mya (Marker & Holmes 2010). This break-up was central in shaping the landscape as the geomorphic processes of erosion, deposition, rifting and warping over time contributed to the contemporary setting of this region.

The Cape Fold Mountains, one of the most notable and important geomorphic features in the area, dominate the landscape - located approximately 20 km from the coast and reaching an altitude of up to 1500m (Martin 1968). Quartzitic sandstones and folded quartzites of the Table Mountain Group Sandstone, Witteberg and Bokkeveld Groups are the main constituents of these mountains and

general area (Martin 1968; Marker & Holmes 2010). Towards the coast, the region is principally underlain by Table Mountain Sandstone and Bokkeveld groups with the Precambrian-Cambrian Malmesbury Group and the Kango and Kaaimans Groups further contributing to the underlying geological composition of the area (Marker & Holmes 2010).

Table 2.1. The main geological units and geomorphic events for the Southern Cape (after Marker & Holmes 2010)

Era	Myr	Period	Group/Formation	Geomorphic Event
Cenozoic	0.01	Holocene	Waenhuiskrans Formation	Bredasdorp Group
		Upper Pleistocene	Klein Brak Formation	
		Middle Pleistocene		
		Early Pleistocene		
	1.8			
		Pliocene	Wankoe Formation	
	5			
		Miocene	Coversands	200m uplift
	25			
		Oligocene	Knysna Formation	200m uplift
Mesozoic	36			
		Eocene		
	54			
		Palaeocene		
	65			
		Cretaceous	Grahamstown Formation	Coastal Platform African Planation
	141			
		Jurassic	Uitenhage Group	Gondwana Rifting
Palaeozoic	200			
	230			
		Triassic		Cape Orogeny
		Perimian		
		Carboniferous	Witteberg Group	
		Devonian	Bokkeveld Group	
		Silurian		Cape Supergroup
		Ordovician	Table Mountain Group	
		Cambrian		
	550			
	600-900	Late Precambrian	Cape Granite Suite Kaaimans & Kango Groups	
	700		Malmesbury Group	

The coastline of the Southern Cape is also characterised by a variety of Pleistocene Aeolian deposits, including the Wilderness dune cordons – a series of fossil dune ridges parallel to the coastline (Bateman et al. 2004). These features will be discussed in more detail in section 2.6.3.

2.5. Vegetation of the Southern Cape

Southern Cape vegetation is characterised by Fynbos and Afrotropical forest, with smaller elements of thicket as well as succulent- and Nama Karoo also present along the fringes of the region (Mucina & Rutherford 2006).

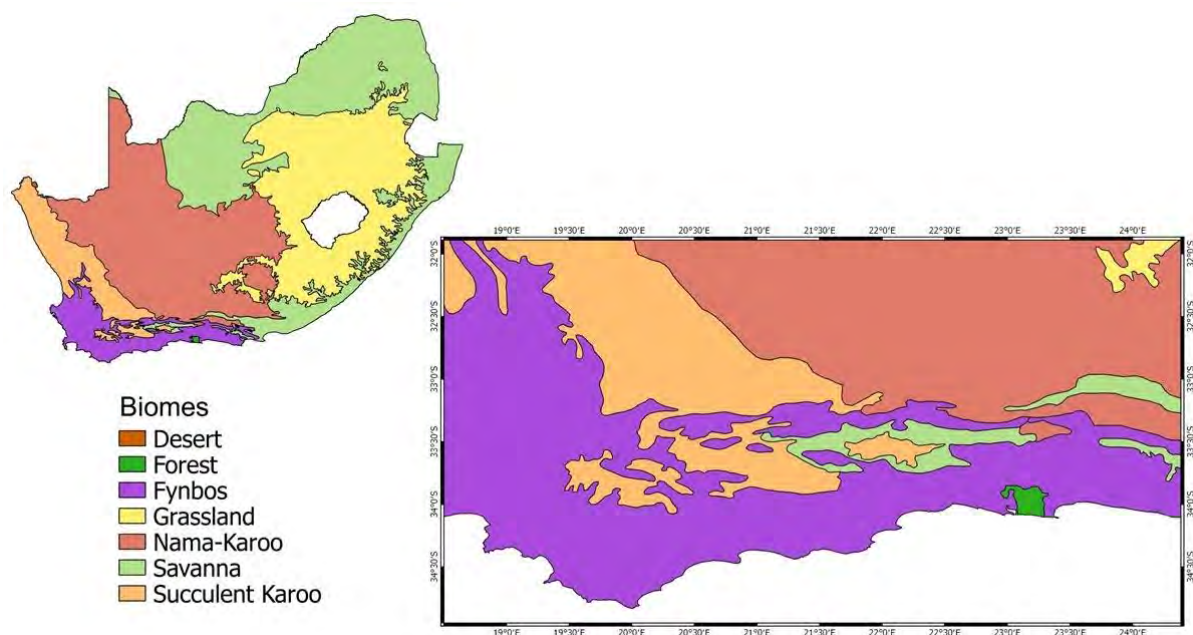


Figure 2.4. Biomes of the southern Cape (Data source: Mucina & Rutherford 2006)

2.5.1. Fynbos

Fynbos vegetation, comprising the largest part of the Cape Floristic Region, denotes an extensive floristic group of assorted evergreen sclerophyllous shrubland (Cowling 1992; Cowling et al. 2004). The Cape Floristic Region covers a 90 000 km² area including the winter rainfall zone in the west and spreading east to include the Southern Cape region (Cowling et al. 2009).

Characteristically, fynbos includes a restiad, ericoid and proteid component (Cowling 1992; Cowling et al. 2004), as set out in Table 2.2.

Table 2.2. Fynbos structural components (adapted from Cowling et al. 2004)

Restiod	Ericoid	Proteoid
Restionaceae or Cape Reed Family	Dominated by Ericaceae or "true heaths"	Almost exclusively Proteaceae
Restionaceae, <i>Tetraria</i> , <i>Leucadendro</i> , <i>Penstachistis</i> , <i>Phylica</i> , <i>Protea</i> , <i>Stoebe</i> , etc.	Ericaceae, Bruniaceae, Fabaceae, Rhamnaceae, Thymelaceae, Restionaceae	<i>Leucadendron</i> spp., <i>Protea</i> spp., Restionaceae, Ericaceae, etc.
Also includes sedges and grasses		

The Fynbos Biome further includes Renosterveld, a transitional vegetation type distinguished by the abundance of *Elytropappus rhinocerotis*, or Renosterbos, as well as other members of the Asteraceae family. Renosterveld vegetation are commonly found on inland valleys and coastal forelands on soils derived from the Karoo Sequence and Malmesbury and Bokkeveld Groups (Low & Rebelo 1996).

A range of factors influence the distribution of fynbos, with climate and soil conditions emphasised as the greatest limiting factors (Cowling 1983a; Mucina & Rutherford 2006) (Cowling 1983; Mucina & Rutherford 2006). The Fynbos Biome generally receives more than 40% of its yearly rainfall during winter (Rutherford & Westfall 1986) with soils having lower available phosphorous and pH levels (Cowling et al. 2004).

Fire plays a major role in fynbos community development - many species depend on fire for regeneration with seeds only released after a fire event (Bond 2004; Privett & Lutzeyer 2010). Fire is a key component in maintaining the borders of the biome in relation to other vegetation types. In the absence of fire, fynbos becomes senescent and vulnerable to invasion by especially thicket and forest elements (Mucina & Rutherford 2006).

2.5.2. Afrot temperate Forest

Southern Cape Afrot temperate Forests, as defined by Mucina and Rutherford (2006), are distributed along the southern coast, between the towns of Mossel Bay and Humansdorp and bordered in the north by the Tsitsikamma-Outeniqua mountain ranges (Geldenhuys 1993). This is the largest forest complex in southern Africa but the distribution is fragmented - numerous large complexes are found on the coastal plain with smaller patches in ravines and mountains (Geldenhuys 1993; Midgley et al. 2004).

Several hypotheses have been put forward to explain the fragmented nature of contemporary forest distribution in the area - the principal of these, as first proposed by Acocks (1953), is that current patches are relicts of previous extensive forests before the large scale fragmentation subsequent to the late Miocene (Geldenhuys 1993). Additionally, it is posited that these patches denote a more recent invasion by forest vegetation (Cowling 1983a; Meadows & Linder 1993), or simply that forest distribution has always been irregular (Midgley et al. 2004)

The contemporary distribution of forests is governed by a variety of factors, including rainfall, topography, soil composition and fire. Rutherford and Westfall (1986) suggest that rainfall is the principal environmental limiting factor with forest development dependant on rainfall upward of 625mm per annum. Geldenhuys (1993) however indicates that, specifically regarding Southern Cape forests, rainfall levels vary, with lower rainfall from 500mm in the vicinity of Great Brak River, up to 1200mm at Storms River, Jonkersberg and Diepwalle.

Additionally, higher altitudes will sustain more forest stands, although smaller in size, with the largest stands present at 250 m.a.m.s.l. (Geldenhuys 1994). Southern Afrotropical forests are typically found on plateaus in the Knysna-Tsitsikamma region as well as steep, south-facing slopes of river catchments, while boulder scree and kloofs provide a refuge from fire (von Maltitz et al. 2003).

With climate largely regulating the broader patterns of forest distribution, fire plays an important role in managing the finer scale patterns, especially relating to forest boundaries (Midgley et al. 2004). Geldenhuys (1994) proposes that berg wind fires are in fact the key determinant of forest distribution as these winds have a desiccating effect on vegetation, increasing the flammability of habitually moist vegetation, such as forests.

In addition to berg wind driven fires, most prevalent during autumn and winter, lightning induced fires tend to occur in summer (Horne 1981). The southern Cape has the highest incidence of these fires, in relation to other forest regions, although the fires are generally small and constrained to mountain ridges (Horne 1981).

2.5.3. The Knysna Afrotropical Region

The Knysna Afrotropical Region (KAR) forms the largest concentration of Southern Cape Afrotropical Forest (Geldenhuys 1993). Scattered among the forest patches several fynbos communities are present – predominantly granite- and shale fynbos (Geldenhuys 1993; Mucina & Rutherford 2006).

Granite fynbos is largely distinguished by the presence of spiny-leaved species, most notably *Cliffortia* spp., while shale fynbos are characterised by larger quantities of herbs and grasses, predominantly found on Kaaimans Group sediments (Mucina & Rutherford 2006).

Geldenhuys (1993) categorised the forest elements present in the KAR as high wet forest, moist forest, moist to dry forest, dry forest and scrub forest. As rainfall increases from the coast to the mountains a sort of gradient of forest type distribution occurs. Dry- and scrub forests are predominantly found along the dry coastal zone while high canopy moist forest inhabit the higher coastal plateau and surrounding hills (van der Merwe 2009). Wet forest elements in turn, are located on mountain slopes and deep, damp ravines supported by an understory of ferns and dense shrubby vegetation (van der Merwe 2009).

Table 2.3-1. Forest categories within the Knysna-Afrotemperate Region [adapted from Geldenhuys (1993) and van der Merwe (2009)]

Common Name	Scientific Name	General description	Location
Moist High Forest			
Stinkwood	<i>Ocotea bullata</i>		
Real yellowwood	<i>Podocarpus latifolius</i>	Canopy up to 22m	
Cape beech	<i>Rapanea melanophloeas</i>		Higher parts of coastal platform,
White alder	<i>Platylophus trifoliatus</i>		poorly drained loamy soils that are wet throughout the year
	<i>Olea capensis</i> subsp		
Ironwood	<i>macrocarpa</i>	Above canopy	
Outeniqua yellowwood	<i>Podocarpus falcatus</i>		
Stinkleaf	<i>Plectranthus fruticosus</i>		
Seven weeks fern	<i>Rhumohra adiantiformis</i>	Ground flora	
Blue sourbulb	<i>Aristea ensifolia</i>		
Wet High Forest			
Stinkwood	<i>Ocotea bullata</i>		Eastern and southern slopes of foothills and higher platform
Red alder	<i>Cunonia capensis</i>	Canopy of between 12 and 20m	bordering mountains, permanently wet shallow soils where rainfall is high
Real yellowwood	<i>Podocarpus latifolius</i>		
Kamassi	<i>Gonioma kamassi</i>	middle sized tree	
Tree fuchsia, notsung	<i>Halleria lucida</i>	small tree	
Wet Scrub Forest			
Red alder	<i>Cunonia capensis</i>		
Stinkwood	<i>Ocotea bullata</i>	Stunted trees, low canopy of between 6 and 10m	High on sheltered mountain slopes and deep ravines, very wet shallow soils
Real yellowwood	<i>Podocarpus latifolius</i>		
Cape beech	<i>Rapanea melanophloeas</i>		
Cape stock rose	<i>Sparmannia africana</i>	Shrub	

Table 2.3-2. Forest categories within the Knysna-Afrotemperate Region [adapted from Geldenhuys (1993) and van der Merwe (2009)]

Common Name	Scientific Name	General description	Location
Dry Scrub Forest	Num-Num	Thorny bush	
	White milkwood		
	Candlewood	Part of low canopy, between 6 and 12m	Warmest sites close to the coast
	Cape saffron		
	Sagewood	Shrub	
Dry High Forest	Candlewood	<i>Pterocelastrus tricuspidatus</i>	
	White pear	<i>Apodytes dimidiata</i>	Moderately warm western and northern slopes of lower coastal platform
	Quar	<i>Psyrax obovata</i> subsp. <i>obovata</i>	
Medium-moist High Forest	Stinkwood	<i>Ocotea bullata</i>	
	Real yellowwood	<i>Podocarpus latifolius</i>	Canopy up to 22m
	Cape beech	<i>Rapanea melanophloeos</i>	
		<i>Olea capensis</i> subsp. <i>macrocarpa</i>	Coastal platform, deep loamy soils kept moist by sufficient rainfall
	Ironwood	<i>Podocarpus falcatus</i>	
	Outeniqua yellowwood	<i>Gonioma kamassi</i>	Intermediate canopy
	Kamassi		Shrub layer in understorey
	Black witch-hazel	<i>Trichocladus crinitus</i>	

2.5.4. The Fynbos-Forest Boundary and the Significance of Fire

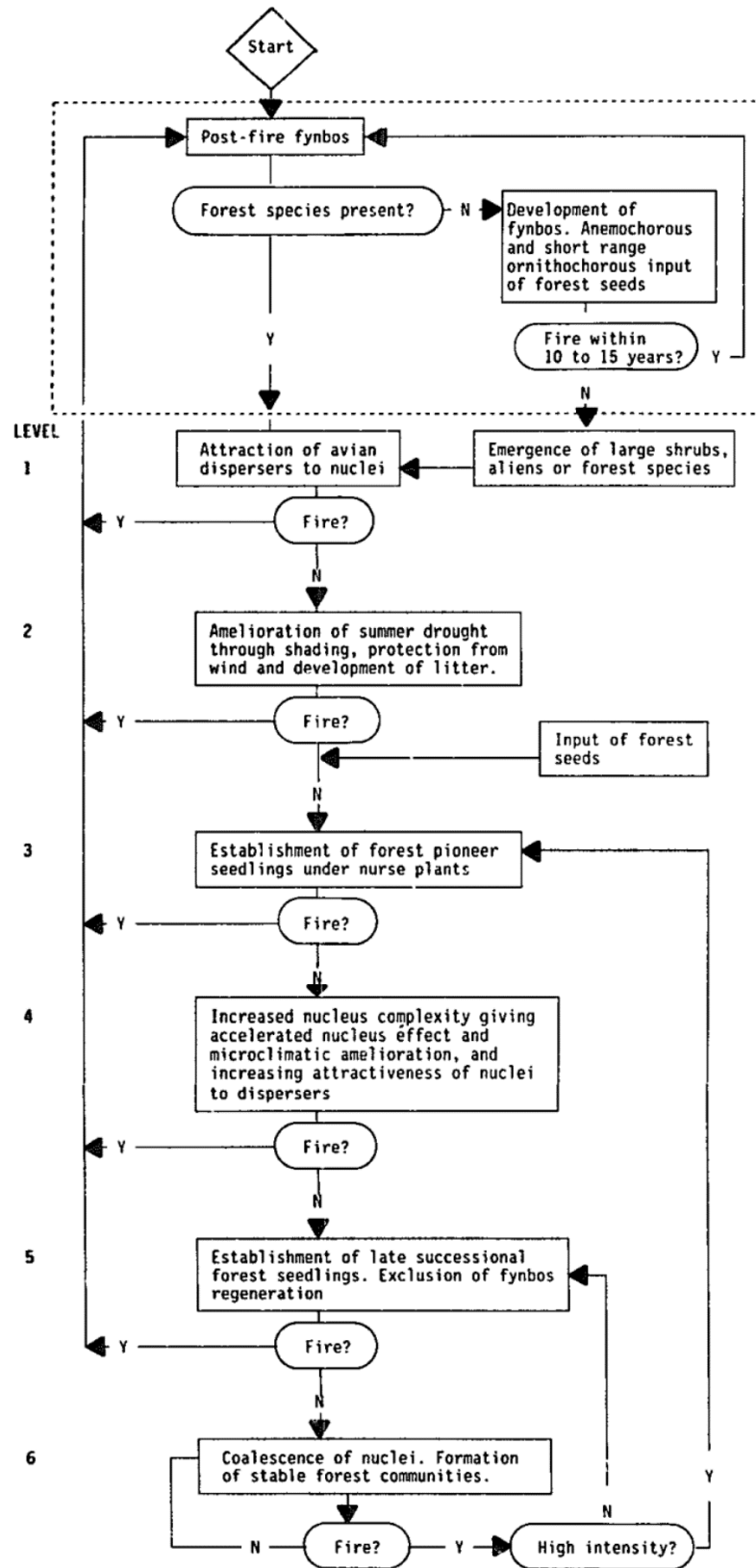


Figure 2.5. The conceptual model displaying the influence of fire on the succession of forest and fynbos communities (from Manders & Richardson 1992)

As mentioned previously, fire plays a pivotal role with regard to fynbos community development and distribution as well as the composition and distribution of forests. Furthermore, fire is essential in maintaining the fynbos-forest boundaries (Cowling 1992; Manders & Richardson 1992; Midgley et al. 2004).

Manders and Richardson (1992) formulated a conceptual model describing these dynamics – this is displayed in Figure 2.7. From this model it is proposed that, with the exclusion of fire in the ecosystem, over time, forests will impinge and progressively replace the surrounding fynbos communities.

2.5.5. Humans vs Forest: A short history of colonisation and forest exploitation

The rapid utilisation, and abuse, of forests started in the 1770's with the arrival of European settlers in the region (Phillips 1931; Geldenhuys 1991; van der Merwe 2009). After extensive exploration of the region, two woodcutter's post were established; at Swartrivier, near George, in 1776 and at Plettenberg Bay in 1787 (Phillips 1931; Geldenhuys 1991; van der Merwe 2009).

In addition to timber harvesting, large forest patches were cleared for agriculture (Geldenhuys 1991; van der Merwe 2009). As a result of the destructive harvesting practices, the Government closed the forests in 1846, selling off large formerly forested areas of land to the public (Phillips 1931; Geldenhuys 1991; van der Merwe 2009).

Unfortunately, most of the forests were reopened for harvesting again in 1856 due to a timber shortage – this time a cyclical harvesting system was however employed (Phillips 1931; Geldenhuys 1991; van der Merwe 2009). A "Conservator of Forests" was appointed in 1874 imposing additional restrictions on timber harvesting (Phillips 1931; Geldenhuys 1991; van der Merwe 2009). Following this, the first significant progress in forest conservation was made by the appointment of a French forestry officer as "Superintendent of Woods and Forests" in 1880, with the Cape Forest Act ratified in 1888 (Phillips 1931; Geldenhuys 1991; van der Merwe 2009).

Exploitation of the forests persisted however, driving the Government to, again, restrict all harvesting practices in 1939 (Phillips 1931; Geldenhuys 1991; van der Merwe 2009). Finally, in 1964, the Indigenous Forest Management Station (IFMS) was established at Saasveld (Phillips 1931; Geldenhuys 1991; van der Merwe 2009). Following this, the forests were reopened to the timber industry in 1965 – under the strict control of the IFMS (Phillips 1931; Geldenhuys 1991; van der Merwe 2009).

2.6. The Wilderness Embayment

Along the South coast, in the area extending from the town of Wilderness to the Goukamma River, the Wilderness-Knysna embayment comprises a 32km section of coastal lakes (Martin 1956). These lakes are separated from the coastline by 200m high sand dunes, and essentially lie in a valley, backed by coastal mountains reaching an altitude of 1200m (Martin 1959).



Figure 2.6 The Wilderness Lakes – Eilandvlei, Langvlei, Rondevlei, Swartvlei and Groenvlei

The Wilderness Lake System, contained within the Wilderness Embayment, consists of Eilandvlei, Bo-Langvlei (also called Langvlei), Rondevlei, Swartvlei and Groenvlei. Eilandvlei, Langvlei and Rondevlei forms part of the Touw River system which has been selected as a ‘Wetland of International Importance’ in 1991 through the Ramsar Convention (Russel et al. 2010), with the Serpentine River connecting Eilandvlei to Langvlei and Rondevlei. Swartvlei is fed by the Karatara, Hoogekraal and Diep rivers while the freshwater input to Groenvlei comes from seepage from the surrounding barrier dunes (Allanson & Whitfield 1983; Russell et al. 2010; Russell et al. 2012).

2.6.1. Local Climate

The contemporary climate of the Wilderness region can be described as temperate with rainfall at a mean annual average ranging from 800 to 1 000mm (Allanson & Whitfield 1983; Russel et al. 2012). This area has no distinct dry season, or marked seasonal rainfall pattern (Martin 1959) but slight increases in rainfall are however experienced from March to April and September to October (Allanson & Whitfield 1983; Russel et al. 2012).

Throughout the year, south-west winds generally prevail, although during winter the wind is predominantly northerly to north-westerly offshore, while in summer this changes to be largely southerly to south-easterly onshore (Deacon & Lancaster 1988). This area is also characterised by the incidence of bergwinds, particularly during winter months, resulting in abnormally high temperatures. Other than this, temperatures are consistently mild throughout the year averaging between 13 and 20 °C annually, with mean daily temperatures during summer ranging from 15 to 25°C and during winter from 7 to 19°C (Allanson & Whitfield 1983).

2.6.2. Local Vegetation

Vegetation in the Wilderness region comprise of a mosaic of fynbos and thicket elements with pockets of Knysna Afromontane forest also present (Cowling & Heijinis 2001). Eleven vegetation types, as described by Mucina and Rutherford (2006) characterise the region, including Southern Afrotemperate Forest, numerous Fynbos types and Azonal vegetation, most prominent along the coast and lake fringes. The location and characteristic species of these vegetation types are set out below in Table 2.4.

Table 2.4. Typical vegetation types of the Wilderness Embayment [adapted from Mucina & Rutherford (2006) and Russel et al. (2012)]

Eastern Coastal Shale Band Vegetation (FFb6)		
Proteaceae	<i>Protea neriifolia</i> <i>Leucadendron eucalyptifolium</i> <i>Leucadendron saliganum</i> <i>Leucospermum cuneiforme</i>	Primarily along the Outeniqua mountains

Southern Afrotemperate Forest (FOz1)		
Podocarpaceae	<i>Afrocarpus falcatus</i> <i>Podocarpus latifolius</i>	Generally limited to river valleys, plains and pediment. Absent from coastal plain
Oleaceae	<i>Olea capensis</i> spp. <i>Macrophylla</i>	
Stlbaceae	<i>Nuxia floribunda</i>	
Celastraceae	<i>Cassine peragua</i>	
Meliaceae	<i>Ekebergia capensis</i>	
Cunoniaceae	<i>Cunonia capensis</i>	
Cornaceae	<i>Curtisia dentata</i>	
Anacardiaceae	<i>Rhus chirindensis</i>	
Myrsinaceae	<i>Rapanea melanophloeos</i>	
Hamamelidaceae	<i>Trichocladus crinitus</i>	
Apocynaceae	<i>Carissa bispinosa</i>	
Vitaceae	<i>Rhoicissus tomentosa</i>	
Ranunculaceae	<i>Clematis brachiata</i>	
Pteridaceae	<i>Pteris dentata</i>	
Dennstaedtiaceae	<i>Hypolepis sparsisora</i>	
Dryopteridaceae	<i>Polystichum lucidum</i>	
Blechnaceae	<i>Blechnum australe</i>	

Southern Coastal Forest (FOz6)		
Sapotaceae	<i>Sideroxylon inerme</i>	Limited distribution within the Garden Route National Park and Goukamma Nature Reserve
Anacardiaceae	<i>Rhus chirindensis</i>	
Celastraceae	<i>Gymnosporia buxifolia</i> <i>Myrtroxylon aethiopicum</i>	
Hamamelidaceae	<i>Trichocladus crinitus</i>	
Ulmaceae	<i>Celtis africana</i>	
Menispermaceae	<i>Cissampelos torulosa</i>	
Apocynaceae	<i>Cynanchum obtusifolium</i> <i>Carissa bispinosa</i>	
Oleaceae	<i>Olea exasperata</i>	
Anacardiaceae	<i>Rhus glauca</i>	
Fabaceae	<i>Acacia cyclops</i> (Rooikrans) <i>Acacia saligna</i> (Port Jackson)	

Southern Cape Dune Fynbos (FFd11)		
Rutaceae	<i>Agathosma apiculata</i>	Intermittent distribution along the Touws River estuary to the east. Narrow belt along the coast
Rhamnaceae	<i>Phylica spp.</i>	
Asteraceae	<i>Metalasia muricata</i>	
Polygalaceae	<i>Muraltia squarrosa</i>	
Cyperaceae	<i>Ficina spp</i>	
Restionaceae	<i>Ischyrolepis</i>	
Sapotaceae	<i>Sideroxylon inerme</i>	
Asteraceae	<i>Tarchonantus littorals</i>	
Celastraceae	<i>Pterocelastrus rostratus</i>	
Lamiaceae	<i>Salvia africana-lutea</i>	
Anacardiaceae	<i>Rhus glauca</i>	
Oleaceae	<i>Olea exasperata</i>	
Fabaceae	<i>Acacia cyclops</i> (Rooikrans)	
	<i>Acacia saligna</i> (Port Jackson)	

Garden Route Shale Fynbos (FFh9)		
Proteaceae	<i>Protea spp.</i>	Discretely dispersed along the coastal plateau
	<i>Leucadendron eucalyptifolium</i>	
	<i>Leucadendron salignum</i>	
Asteraceae	<i>Elytropappus rhinocerotis</i>	
	<i>Metalasia densa</i>	
Rhamnaceae	<i>Phylica axillaries</i>	
	<i>Phylica pinea</i>	
Dennstaedtiaceae	<i>Pteridium aquilinum</i>	
Ericaceae	<i>Erica hispidula</i>	
Crassulaceae	<i>Crassula orbicularis</i>	
Poaceae	<i>Aristida junciformis</i>	
	<i>Brachiaria serrata</i>	
	<i>Eragrostis capensis</i>	
	<i>Themeda triandra</i>	
Restionaceae	<i>Ischyrolepis gaudichaudiana</i>	
	<i>Ischyrolepis sieberi</i>	
	<i>Elegia juncea</i>	
	<i>Restio triticeus</i>	

Garden Route Granite Fynbos (FFg5)		
Santalaceae	<i>Osyris compressa</i>	West of the Kaaimans River mouth along the coast
Thymelaceae	<i>Passerina corymbosa</i>	
Proteaceae	<i>Protea</i> spp.	
	<i>Leucadendron saliganum</i>	
	<i>Mimetes cucullatus</i>	
Rosaceae	<i>Cliffortia serpyllifolia</i>	
Ericaceae	<i>Erica discolor</i>	
	<i>Erica canaliculata</i>	
	<i>Erica peltata</i>	
Rhamnaceae	<i>Phyllica confusa</i>	
Rutaceae	<i>Agathosma ovata</i>	
Asteraceae	<i>Eriocephalus africanus</i>	
Geraniaceae	<i>Pelargonium fruticosum</i>	
Mesembryanthemaceae	<i>Lampranthus sociorum</i>	
Poaceae	<i>Eragrostis capensis</i>	
	<i>Heteropogon contortus</i>	
	<i>Pentaschistis eriostoma</i>	
	<i>Themeda triandra</i>	
Cyperaceae	<i>Tetraria cuspidata</i>	
	<i>Ficinia nigrescens</i>	
Restionaceae	<i>Restio triticeus</i>	

Knysna Sand Fynbos (FFd10)		
Thymelaceae	<i>Passerina rigida</i>	Northern side of lakes, sparse
Proteaceae	<i>Leucadendron saliganum</i>	
Polygalaceae	<i>Muraltia squarrosa</i>	
Ericaceae	<i>Erica cuvifolia</i>	
Cupressaceae	<i>Widdringtonia nodiflora</i>	
Asteraceae	<i>Metalasia densa</i>	
Fabaceae	<i>Tephrosia capensis</i>	

Southern Outeniqua Sandstone Fynbos (FFs19)		
Sapindaceae	<i>Dodonaea angustifolia</i>	From east to west along the (south-facing) slopes of the Outeniqua Mountains.
Cupressaceae	<i>Widdringtonia nodiflora</i>	
Stilbaceae	<i>Halleria lucida</i>	
Proteaceae	<i>Protea neriifolia</i>	
	<i>Protea repens</i>	
	<i>Protea mundii</i>	
	<i>Leucadendron eucalyptifolium</i>	
Asteraceae	<i>Metalasia densa</i>	
	<i>Stoebe alopecuroides</i>	
	<i>Helichrysum felinum</i>	
	<i>Osteospermum monilifera</i>	
	<i>Relhania calycina</i>	
	<i>Euryops pinnatipartitus</i>	
Anacardiaceae	<i>Laurophyllus capensis</i>	
Bruniaceae	<i>Brunia nodiflora</i>	
	<i>Berzelia intermedia</i>	
Ericaceae	<i>Erica</i> spp.	
Proteaceae	<i>Leucadendron salignum</i>	
	<i>Protea cynaroides</i>	
Rutaceae	<i>Acmadenia</i> spp.	
Rosaceae	<i>Cliffortia ilicifolia</i>	
Fabaceae	<i>Otholobium carnum</i>	
Apiaceae/Araliaceae	<i>Centella affinis</i>	
Apiaceae	<i>Anginon difforme</i>	
	<i>Geissorhiza</i> spp.	
Lanariaceae	<i>Lanaria lanata</i>	
Iridaceae	<i>Tritoniopsis caffra</i>	
	<i>Watsonia fourcadei</i>	
Poaceae	<i>Ehrharta dura</i>	
	<i>Ehrharta rupestris</i>	
	<i>Pentameris distichophylla</i>	
	<i>Merxmüllera decora</i>	
	<i>Pentaschistis</i> spp.	
Cyperaceae	<i>Tetraria</i> spp	
Restionaceae	<i>Cannamois</i> spp.	
	<i>Elegia</i> spp.	
	<i>Restio</i> spp.	
	<i>Thamnochortus</i> spp.	
Pinaceae	<i>Pinus pinaster</i> (Pine)	
Proteaceae	<i>Hakea sericea</i> (Silky Hakea)	

Garden Route Shale Fynbos (FFh9)		
Proteaceae	<i>Protea</i> spp. <i>Leucadendron eucalyptifolium</i> <i>Leucadendron salignum</i>	Dispersed along the coastal plateau
Asteraceae	<i>Elytropappus rhinocerotis</i> <i>Metalasia densa</i>	
Rhamnaceae	<i>Phylica axillaries</i> <i>Phylica pinea</i>	
Dennstaedtiaceae	<i>Pteridium aquilinum</i>	
Ericaceae	<i>Erica hispidula</i>	
Crassulaceae	<i>Crassula orbicularis</i>	
Poaceae	<i>Aristida junciformis</i> <i>Brachiaria serrata</i> <i>Eragrostis capensis</i> <i>Themeda triandra</i>	
Restionaceae	<i>Ischyrolepis gaudichaudiana</i> <i>Ischyrolepis sieberi</i> <i>Elegia juncea</i> <i>Restio triticeus</i>	

Cape Seashore Vegetation (AZd3)		
Asteraceae	<i>Arctotheca populifolia</i> <i>Gazania rigens</i> <i>Senecio elegans</i>	Narrow belt along the seashore. Beaches, coastal dunes and cliffs
Geraniaceae	<i>Pelargonium capitatum</i>	
Solanaceae	<i>Solanum africanum</i>	
Apocynaceae	<i>Cynachum ellipticum</i> <i>Cynachum obtusifolium</i>	
Aizoaceae	<i>Carpobrotus edulis</i> <i>Carpobrotus acinaciformis</i> <i>Tetragonia decumbens</i> <i>Drosanthemum</i> spp.	
Apiaceae	<i>Dasispermum suffruticosum</i> <i>Ehrharta villosa</i>	
Poaceae	<i>Sporobolus virginicus</i> <i>Thinopyrum distichum</i>	

Cape Estuarine Salt Marshes (AZe2)		
Ruppiaceae	<i>Ruppia maritima</i> <i>Ruppia cirrhosum</i>	Typically restricted to the Knysna and Swartvlei estuaries
Zosteraceae	<i>Zostera capensis</i>	
Chenopodiaceae	<i>Sarcocornia perennis</i> <i>Sarcocornia capensis</i> <i>Sarcocornia pillansii</i>	
Amaranthaceae	<i>Chenolea diffusa</i>	
Plantaginaceae	<i>Plantago crassifolia</i>	
Juncaceae	<i>Juncus kraussi</i>	
Poaceae	<i>Sporobolus virginicus</i> <i>Stenotaphrum secundatum</i>	

Cape Lowland Freshwater Wetlands (AZf1)		
Typha	<i>Typha capensis</i>	Bordering the lakes. Behind dune cordons
Juncaceae	<i>Juncus rigidus</i> <i>Juncus krassii</i>	
Cyperaceae	<i>Schoenoplectus scirpoides</i> <i>Cladium mariscus</i> <i>Ficinia nodosa</i> <i>Cyperus textilis</i> <i>Cyperus sphaerospermum</i> <i>Cyperus thunbergii</i> <i>Pycnus nitidus</i>	
Poaceae	<i>Stenotaphrum secundatum</i> <i>Phragmites australis</i> <i>Pennisetum macrourum</i> <i>Paspalum vaginatum</i>	
Aponogetonaceae	<i>Aponogeton distachyos</i> <i>Aponogeton junceus</i>	
Menyanthaceae	<i>Nymphoides indica</i>	
Ceratophyllaceae	<i>Ceratophyllum demersum</i>	
Potamogetonaceae	<i>Potamogeton pectinatus</i>	
Ruppiaceae	<i>Ruppia maritima</i>	

In the immediate vicinity of the Wilderness lakes, in the low lying floodplain between the normal water level and the 5m contour a.m.s.l, vegetation can be classified into two general categories - semi-aquatic and terrestrial vegetation (Allanson & Whitfield 1983).

Semi-aquatic vegetation, underlain by poorly drained soils rich in alluvium, include bulrushes (*Typha latifolia*), reeds (*Phragmites australis*) and sedges (*Scirpus littoralis*) (Allanson & Whitfield 1983; Russel

et al. 2012). The area inundated during extremely high water levels represents an intermediary zone between the semi-aquatic and terrestrial environments and is covered by rushes such as *Juncus kraussi* (Allanson & Whitfield 1983; Russel et al. 2012).

Terrestrial vegetation is presented by characteristic coastal maccia fynbos, which include *Carpobrotus*, *Delosperma*, *Chironia*, Restionaceae and ericoid-leaved plants such as *Metalasia muricata* (Allanson & Whitfield 1983; Russel et al. 2012). These grow on humus-deficient, well drained sandy soils (Allanson & Whitfield 1983; Russel et al. 2012).

2.6.3. Geology and Geomorphology: Dune Cordons and Lake Evolution

Geomorphologically, the Wilderness region can be described by three features: the Outeniqua Mountains in the North, the coastal plateau and the coastal embayment.

The Outeniqua Mountains are primarily composed of quartzites and sandstones of the Table Mountain Group with a maximum altitude of 1600 m.a.m.s.l. (Martin 1962; Russel et al. 2012). The coastal plateau represents the old sea floor with Tertiary origin, comprised of Table Mountain sandstones, pre-Cape granite and Kaaiman Group sediments (Martin 1962; Illenberger 1996). The coastal embayment dates from the Pleistocene and predominantly consists of dune deposits forming the characteristic dune cordons (Bateman et al. 2004; Bateman et al. 2011).

The dune cordons, running parallel along the coast, are identified as the landward, middle and seaward barriers (Illenberger 1996; Bateman et al. 2004; Bateman et al. 2011). The landward barrier is the oldest, located east of Knysna, and extends up to 280 m.a.m.s.l. (Illenberger 1996; Bateman et al. 2011; Russel et al. 2012). The middle barrier ranges to a height of 210 m.a.m.s.l., spanning the area between Langvlei and the Goukamma River (Bateman et al. 2011). The seaward cordon, the youngest of the barriers, stretches from Wilderness to Brenton-on-Sea (Bateman et al. 2011; Russel et al. 2012).

Martin (1962) and Illenberger (1996) postulated that these dune cordons were formed during the sea level high stands of the Pleistocene interglacials and attributes the development of these dunes, and the lakes separating them, to sea level oscillations of the Pleistocene. Bateman et al. (2011) agrees, adding that greater aeolian activity, and thus dune building, was a result of the prevailing westerlies. In addition to Pleistocene dune building phases, increased aeolian activity was noted during the late Holocene from c. 3.7 to 2.4 ka and again between c. 1.7 and 0.6 ka (Bateman et al. 2011).

It is proposed that the Wilderness lakes formed between 67 000 and c. 200 000 BP, as a result of flooding of the river valleys and the subsequent development of the dune cordons, which cut off the

ivers (Bateman et al. 2004). Martin (1962) suggests that three stages of “submergence and emergence”, the last during the Holocene, resulted in the formation of these lakes, with these stages linked to sea transgressions and regressions.



Table 2.5. The evolution of the Wilderness Embayment (after Martin 1962)

Time	Sea Level	Coastal Evolution	Lake Evolution
Recent C14 date: 52 AD ± 60 years	Stationary or falling slightly	River mouth bars, existing marshes of Swartvlei estuary. Most recent dune surfaces	Variable - tendency to freshwater conditions and fen formation Groenvlei Isolated
C14 date: 4913 BC ± 160 years	Maximum Possibly +1.5 m	2.5m terrace of Swartvlei. Development of lower Gericke Point platform. Dunes migrating inland	Brackish phase in lakes invaded by sea Marine incursion
	Rising	Shore retrograding; erosion of upper Gericke Point platform. New dune system initiated (?)	Drainage impeded in lateral valleys, lower lake mud and peat
	(?) Low	Dune ridge IV, erosion of margin of Sedgefield terrace	Dune valleys (?) dry. Origin of Eilandvlei. Rondevlei originates by wind deflation
		Terrigenous sand deposited above beach and levelled to +7m	"30m" dune between Groenvlei and Sedgefield
Pleistocene	High (c. +5m)	5m shell beach or bank formed across end of Sedgefield-Groenvlei valley. Cutting of main Gericke Point platform	Swartvlei and Groenvlei flooded but partially isolated by beach formation at Sedgefield
	Very low (below - 30m)	Incision of lower river courses	(?) Deepening of lateral valleys by erosion
	Sinking slowly	Origin of dune ridges I, II and III, oldest parts of IIIA	Origin of lakes (except Eilandvlei and Rondevlei) probably as tidal lagoons behind bars, or diverted river courses
	High, not less than 7 to 8m	Origin of ancient 7m beach seen at Gericke Point and Eilandvlei	

2.7. Conclusion

This chapter presented the contemporary Southern Cape region in terms of its climatic, ecological and geological characteristics, while further giving insight into the origin and development of the Wilderness Embayment, and Eilandvlei.

As the Southern Cape is located within the year round rainfall zone, the following chapter provides a review of selected palaeoenvironmental studies from all three rainfall zones.

3. Late Holocene Environmental and Climate Change in South Africa

3.1. Introduction

As the Southern Cape is located within the year round rainfall zone, the climate of the region is influenced by both the temperate westerlies and tropical easterlies from the winter- and summer rainfall zones, respectively. As such, an understanding of Late Holocene environmental and climate change in the winter- and summer rainfall zones of South Africa are needed.

Accordingly, this chapter provides a review of this most prominent palaeoenvironmental records from the winter- and summer rainfall zones, followed by a review of the available information for the year round rainfall zone – the study area in question.

3.2. Defining the Holocene

The onset of the Holocene, the most recent stratigraphic epoch, is widely regarded to be 11 500 cal yr BP (Mayewski et al. 2004). This marks the end of the Younger Dryas, an unusual cold event during the late Pleistocene (Walker et al. 2008), and the resumption of postglacial warming.

With regard to climate, the Holocene is generally divided into two parts. The first is related to conditions associated with the so called Holocene Altithermal (or ‘optimum’) - a global phenomenon spanning the period c. 8 000 to 6 000 cal yr BP characterised by warmer and generally drier environments. During the latter half of the Holocene, conditions evolved to those approximating the present.

More formally, Walker et al (2012) suggests the division of the Holocene into three parts – the early, mid and late Holocene – with the boundaries between these punctuated by noteworthy climatic events. As such, the early-mid Holocene boundary is placed at 8.2 ka BP, marking the ‘8.2 event’ – a brief cooling event before the onset of Altithermal conditions. The boundary between the mid and late Holocene is then suggested to be at 4.2 ka BP, associated with extensive aridity and warming in the Southern Hemisphere (Mayewski et al. 2004).

3.3. Late Holocene Palaeoenvironments of South Africa

3.3.1. The Winter Rainfall Zone (WRZ)

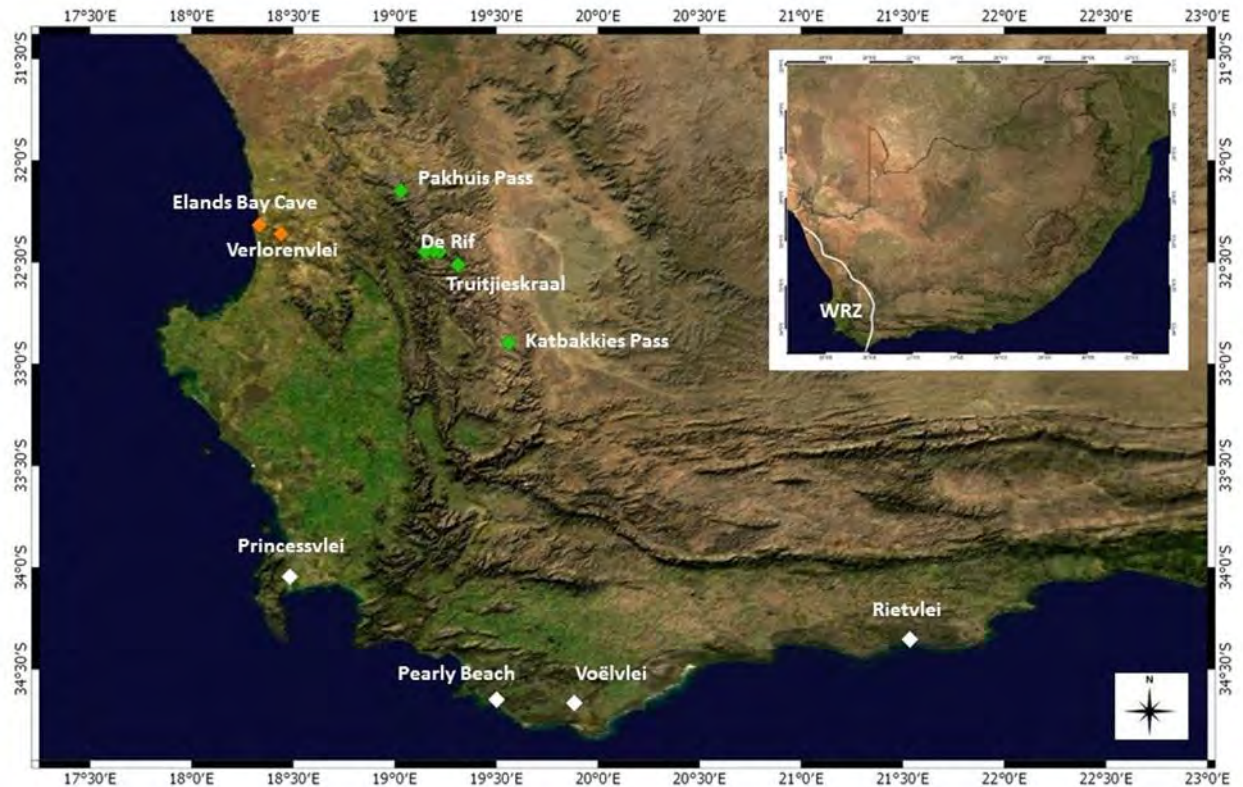


Figure 3.1. The locations of the sites from the winter rainfall zone discussed in the text

The West Coast

The West coast of South Africa has enjoyed a considerable amount of attention with regard to palaeoecological and palaeoclimatological studies, specifically the coastal lake Verlorenvlei situated near the town of Elands Bay (i.e. Meadows et al. 1996; Meadows & Baxter 2001; Baxter & Meadows 1999; Stager et al. 2012).

An outstanding feature from the records from this area, Verlorenvlei especially, is that the largest contributor to, or driver behind, vegetation change is fluctuating sea level (Meadows et al. 1996; Meadows & Baxter 2001) with its influence apparent in the records up until the last 300 years when European colonisation started to leave its distinct mark on the surrounding environment (Meadows et al. 1996; Meadows & Baxter 2001; Baxter & Meadows 1994; Stager et al. 2012).

At the onset of the late Holocene, c. 4 200 cal yr BP, sea levels were at a highstand of 2.8 m above present levels (Miller et al. 1995). Evidence from the Langebaan Lagoon corroborates this suggesting higher sea levels from c. 6 000 to 4 000 cal yr BP (Compton 2001). Likewise, the sea level record from Grootdrift, a coring site in Verlorenvlei, shows a +3 m.a.m.s.l. transgression between c. 5 000 and 4 000 cal yr BP along the West Coast (Baxter & Meadows 1999).

The persistence of an estuarine environment during this period is also evident in the Klaarfontein record in the continual presence of the halophytic taxa *Chenopodiaceae* and *Limonium*-type (Meadows & Baxter 2001). Meadows and Baxter (2001) indicates that, until c. 3 700 cal yr BP, a definite marine influence was present in the vlei, albeit to varying degrees.

From around c. 4 000 cal yr BP charcoal evidence from Elands Bay Cave points to xeric thicket taxa being a dominant component of the vegetation composition suggesting warm and dry conditions (Cartwright & Parkington 1997; Cowling et al. 1999). Meadows et al. (1996) however propose that drier and warmer conditions prevailed during the mid-Holocene, followed by an increase in moisture availability. Both the Klaarfontein and Grootdrift pollen records indicate this, showing greater moisture availability from c. 4 000 to 2 000 cal yr BP (Baxter & Meadows 1999; Meadows & Baxter 2001).

A decline in sea levels to -1 m.a.m.s.l. is suggested at Verlorenvlei between c. 3 700 and 2 800 cal yr BP (Baxter & Meadows 1999). The timing of this regression coincides with the onset of a depositional hiatus in the Grootdrift core with Meadows and Baxter (2001) mentioning that these lower sea levels may have also influenced sediment accumulation at the Klaarfontein site, possibly affecting the pollen record. The sea surface temperature (SST) record from Cohen et al. (1992), derived from aragonite and $\delta^{18}\text{O}$ values in mollusc shells, shows lower SST's between c. 4 000 to 2 000 cal yr BP in the Benguela Upwelling System. Compton (2001) further reports lower sea levels, around -1 to -2 m.a.m.s.l., at Langebaan from c. 2 500 to 1 800 cal yr BP.

The Klaarfontein pollen record points to the return of freshwater conditions to the vlei c. 1 800 cal yr BP, with the occurrence of *Myriophyllum* sp. and *Apongeton* sp. as well as the saline intolerant reeds *Cyperus* sp. and *Typha* sp. (Meadows & Baxter 2001). Indications are thus that moister conditions predominated during this period with the establishment of various fynbos communities as well as the presence of arboreal taxa elements such as *Olea* and *Podocarpus*. More saline/estuarine conditions, and perhaps a drier environment is inferred from the Klaarfontein record after c. 1 800 cal yr BP, with the reoccurrence of *Chenopodiaceae* and *Mesembryanthemum*, as well as a decline in *Typha*, pointing to a possible increase in sea levels. The Grootdrift sea level record does show a slight transgression at

1 350 cal yr BP (Baxter & Meadows 1999) with Compton (2001) suggesting a sea level increase of between +0.5 and +1 m.a.m.s.l. around c. 1 300 cal yr BP at Langebaan.

The high resolution diatom record from Verlorenvlei, shows relatively high rainfall for the period c. 1 400 to 1 200 cal yr BP (Stager et al. 2012). It is interesting to note that the Klaarfontein pollen record makes no specific inferences regarding this period. Evidence from Elands Bay Cave does however indicate a generally stable period regarding vegetation development between c. 1 400 to 320 cal yr BP, with xeric thicket and Asteraceous shrubland still being the dominant components in the assemblage (Cowling et al. 1999).

The development of more swampy conditions from c. 1100 to 960 cal yr BP, at Verlorenvlei, are inferred from the substantial percentage of epiphytic diatoms in the record (Stager et al. 2012). Stager et al. (2012) postulates that the presence of these epiphytic diatoms can be an effect of decreasing lake levels with precipitation registering a decline until c. 950 cal yr BP. Consequently, indications are that generally drier conditions prevailed for the duration of the Medieval Warm Period – or Medieval Climate Anomaly, c. 1 050 to 550 cal yr BP, in this area (Stager et al. 2012).

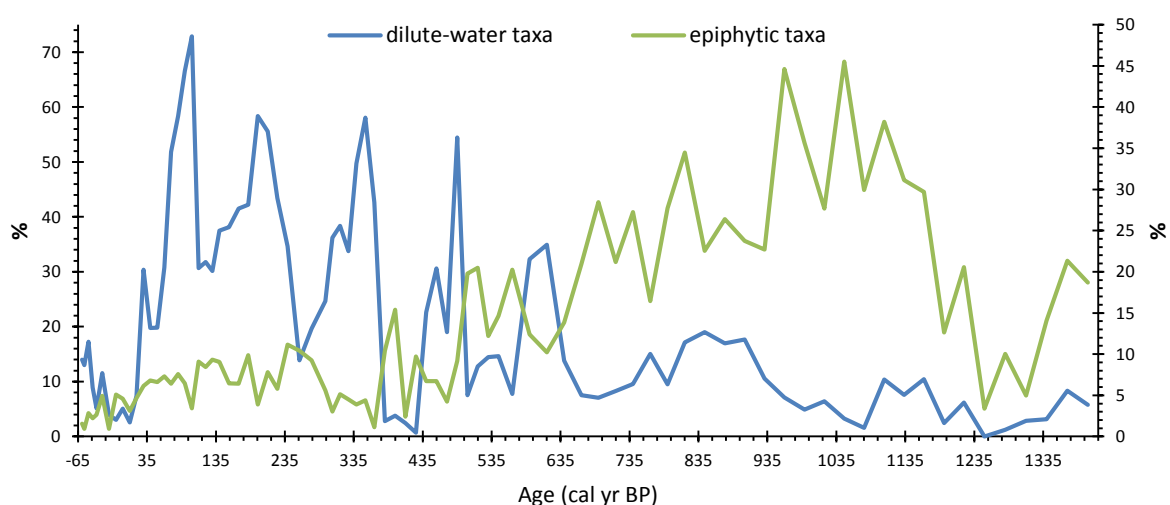


Figure 3.2. c.1 400 year record of diatom abundance at Verlorenvlei (after Stager et al. 2012)

The last 700 years are concluded to be a markedly wet period along the West Coast, with a significant increase in precipitation, as evidenced by the presence of dilute diatom taxa in the record (Stager et al. 2012). Evidence presented by Baxter and Meadows (1994) also supports an environment with increased moisture availability for the period predating the arrival of European settlers in the area. Furthermore, SST's along the West Coast were generally cooler until c. 500 cal yr BP, with a marked

drop in temperatures of between 1 and 2°C at c. 500 cal yr BP (Cohen et al. 1992) while a sea level lowstand of approximately -0.5 to -1 m.a.m.s.l. is recorded at Langebaan around c. 700 cal yr BP, with a transgression exhibited from this point to c. 400 cal yr BP (Compton 2001). This may suggest that the Little Ice Age (LIA), along the West Coast, was generally cool and wet. Increased sediment delivery is also evident at Verlorenvlei from around c. 550 cal yr BP suggestive of variable climatic conditions during the LIA (Stager et al. 2012)

After the depositional hiatus, the Grootdrift record resumes c. 300 cal yr BP with all signs pointing to open fresh water conditions prevailing in the area - the aquatics *Typha* sp., *Apongeton* sp. and *Myriophyllum* sp. are abundant in the pollen spectrum (Baxter & Meadows 1999). This is possibly a result of increased rainfall as the diatom record show increased rainfall pulses at c. 470, 330 and 200 cal yr BP (Stager et al. 2012).

From c. 300 cal yr BP, the records become more difficult to interpret as colonisation, and thus exploitation of the environment, was in full effect (Baxter & Meadows 1994; 1999). The impact of this environmental disturbance is especially evident in the vegetation encroachment around the Grootdrift site – fresh water aquatics decline and are replaced by emergent reeds such as Cyperaceae and *Juncus* as agricultural practices expand and Hippopotamus become locally extinct (Baxter & Meadows 1994; 1999). This anthropogenic effect is further evident in the increased presence of Scrophulariaceae and *Stoebe*-type pollen – common disturbance indicators, while Fynbos elements are on the decline (Baxter & Meadows 1994).

The Southwestern Cape Coast

For the purpose of this review, the Southwestern Cape coast includes the area from Cape Town to Stilbaai. Representing a lesser studied region of the WRZ, most of the records emanating from the southwestern coast do include a late Holocene section, but the overall record tends to be fragmented.

Records from Princessvlei, a lake in the Cape Peninsula, provide the most comprehensive view of late Holocene conditions in this area. The pollen record presents a c. 4 200 year record and includes the period of colonial settlement in the region from c. 300 years ago (Neumann et al. 2011). There are however uncertainties regarding the chronology of this record. A new diatom and isotope record for Princessvlei, encompassing the last c. 2 600 years (Kirsten 2014), has become available which might resolve these issues.

A new multiproxy record from Rietvlei, near Stilbaai, indicates a generally warmer and/or drier climate for most of the middle Holocene, c. 6 900 to 3 300 cal yr BP (Quick et al. in prep). Increased seasonal rainfall as well as episodic warmer and drier conditions are evident for the period c. 4 700 to 3 800 cal

yr BP, with peaks in Crassulaceae, *ChenoAm*-type and Aizoaceae pollen in the record (Quick et al. in prep). Moderately dry conditions also prevailed at Princessvlei c. 4 200 to 3 400 cal yr BP as shown by increases in Asteraceae, Aizoaceae and *Crassula*-type pollen (Neumann et al. 2011).

Wetter conditions, inferred from increases in Cyperaceae and *Morella*, are shown at Princessvlei from c. 3 400 to 2 600 cal yr BP (Neumann et al. 2011a), with the Princessvlei $\delta^{18}\text{O}$ record exhibiting enriched values from c. 3 200 to 2 500 cal yr BP, indicating a change towards cooler conditions (Kirsten 2014). Additional evidence of wetter and cooler conditions are colder water temperatures at c. 2 500 cal yr BP and higher lake levels from c. 2 600 to 2 400 cal yr BP (Kirsten 2014).

A trend towards drier conditions are observed in the Princessvlei pollen record from c. 2 600 cal yr BP with aridity peaking at c. 2 000 cal yr BP (Neumann et al. 2011). The diatom sequence records a decline in water levels c. 2 400 to 2 100, substantiating this drier interval although the timing of this event varies by 200 years between two records (Kirsten 2014). Further along the coast, several episodes of increased aridity are similarly experienced on the Agulhas plain from c. 2 800 to 2 600 cal yr BP, inferred from lunette formation at Voëlvlei (Carr et al. 2006 a, b).

A return to wetter conditions is indicated at Princessvlei c. 2 000 cal yr BP with increases in Capraceae, Cyperaceae and Ericaceae (Neumann et al. 2011). The increased presence of *Nymphaea* pollen further suggests deeper water levels and wetter conditions from c. 1 900 to 1 000 cal yr BP (Neumann et al. 2011). The $\delta^{18}\text{O}$ series points to moderately warm conditions between c. 2200 and 1000 cal yr BP, which may have persisted until c. 800 cal yr BP (Kirsten 2014). Kirsten (2014) points out that during this wet interval, moisture availability fluctuates in the region but conditions largely remain wet until c. 1 530 cal yr BP. The diatom record does note a dry event at c. 1 650 cal yr BP, followed by a brief wetter phase until c. 1 530 cal yr BP (Kirsten 2014).

Kirsten (2014) highlights the period c. 1 530 to 1 300 cal yr BP as a 'transitional phase' marked by a distinct change in the diatom community. Oscillations in lake levels are also evident c. 1 400 to 1 200 cal yr BP although this period was generally wetter (Kirsten 2014). It is further emphasised that environmental conditions were somewhat variable at this time, with the increasing influence of the South Atlantic Anticyclone (SAA) possibly the driver behind this instability (Kirsten 2014).

The Rietvlei record shows the dominance of Fynbos in the assemblage, from c. 2 200 to 900 cal yr BP with $\delta^{13}\text{C}$ values at an all-time low c. 2 000 to 1 200 cal yr BP (Quick et al. in prep). This led Quick et al. (in prep) to propose cooler conditions and increased rainfall, linked to the reassertion of westerlies, for the WRZ for this period. Increased moisture availability is likewise reported at Voëlvlei, along the Agulhas plain around c. 2 000 cal yr BP (Carr et al. 2006 a, b).

Constant $\delta^{18}\text{O}$ levels are however observed between c. 1 800 and 1 200 cal yr BP, with considerably lower values noted at c. 1 200 cal yr BP representing warmer climatic conditions (Kirsten 2014). The nominal presence of *M. nummuloides* in the diatom record further supports the onset of drier conditions c. 1 200 cal yr BP, with the dominant influence of the SAA suggested to initiate this dry period (Kirsten 2014). This arid event continues until c. 920 cal yr BP marked by lower lake levels and decreased moisture availability (Kirsten 2014).

From c. 1 000 to 600 cal yr BP climatic and environmental variability is again inferred at Princessvlei, especially from the $\delta^{18}\text{O}$ record (Kirsten 2014). Intermittent moisture availability is also indicated for this period although a general trend towards a wetter environment is evident (Kirsten 2014). Kirsten (2014) further suggests that the timing of this instability relates to the globally recognised climatic event the Medieval Climate Anomaly, or Medieval Warm Period. Environmental variability, and a trend towards drier conditions are observed at Rietvlei from around c. 900 cal yr BP with the onset of a more aseasonal rainfall pattern and increased temperatures (Quick et al. in prep). Greater aridity is similarly shown along the Agulhas plain at c. 700 (Carr et al. 2006 a, b). The Princessvlei diatom record again points to an intermittent phase between c. 700 and 600 cal yr BP, corresponding to the onset of the Little Ice Age, with mostly cooler and wetter conditions persisting until c. 440 cal yr BP (Kirsten 2014).

Both Princessvlei records show a depositional hiatus, the diatom record from c. 440 to 300 cal yr BP, and the pollen record much longer from c. 1 000 to 350 cal yr BP. Kirsten (2014) posits that the dry/arid event marking the division between the two phases of the LIA might explain this sedimentation break. Correspondingly, increased aridity is again evident along the Agulhas plain around c. 450 cal yr BP (Carr et al. 2006 a, b) with drier phase also indicated at Pearly Beach between c. 450 and 320 cal yr BP (Quick 2013). Quick (2013) further notes remarkably high proportions of Proteaceae from c. 400 cal yr BP suggesting that higher temperatures and increased levels of CO_2 may be responsible for the increased presence of woody thicket elements.

At resumption of the Princessvlei pollen record c. 350 cal yr BP, the increasing effect of anthropogenic disturbances on the environment are evident in increased levels of Poaceae as well as the presence of neophytes including *Pinus* and *Zea mays* (Neumann et al. 2011).

The Cederberg Region

The majority of the information emerging from this region results from the analysis of hyrax middens (e.g. Scott & Bousman 1990; Scott & Woodborne 2007 a,b; Meadows et al. 2010; Quick et al. 2011) with a major focus on the Last Glacial Maximum. Most of these records do include a late Holocene section although, this is discontinuous in some cases.

General indications from the pollen records are that the (late) Holocene in the Cederberg area was largely uneventful with regard to vegetation and climate change. Significant variations with relation to precipitation and vegetation are however evident in the Pakhuis Pass record, especially relating to the last 2 500 years. New isotopic and seasonal rainfall records from Katbakkies (KB-3-1; Chase et al. 2015) further suggest that the Holocene in the Cederberg might in fact not have been as stable as previously thought.

Arid conditions seem to prevail in the region from c. 8 000 to 3 000 cal yr BP, indicated by marginally increased proportions of succulents and increased Asteraceae presence in the Truitjies Kraal pollen record (Meadows et al. 2010). The $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ series from Truitjies Kraal further supports the notion of general aridity during this period (Meadows et al. 2010).

Recently released records from the KB-3-1 midden do however imply markedly greater moisture availability from c. 6 900 to 5 600 cal yr BP, and points to increased summer rainfall in the region for this period; a gradual decline in summer rainfall is however evident until c. 4 900 cal yr BP (Chase et al. 2015). The Pakhuis Pass pollen record corroborates increased moisture, although brief, further highlighting the decreased occurrence of succulents c. 5 600 cal yr BP and the persistent presence of Asteraceae until c. 5 000 cal yr BP (Scott & Woodborne 2007a, b). A further prominent change is highlighted in the Pakhuis Pass isotope record between c. 5 600 and 4 900 cal yr BP indicating increased C3 vegetation in the environment as well as the possible weakening impact of summer rainfall (Scott & Woodborne 2007a, b). Chase et al. (2015) similarly note a change in precipitation seasonality at Katbakkies for the period c. 5 600 to 4 700 cal yr BP, with increased winter rainfall due to the equatorward movement of the westerly storm track, and a concurrent reduction in rainfall in summer leading to a moderately drier environment.

Brief intervals of increased humidity are observed at De Rif (DR2010; Valsecchi et al. 2013) at c. 4 600 and c. 3 200 cal yr BP with the DR-1 record (Quick 2009; Quick et al. 2011), also from De Rif, indicating a reasonably moist environment from c. 3 500 to 2 500 cal yr BP as evidenced by the maximum proportions of Cyperaceae pollen in the record. Similarly, a relatively humid period is evident at Katbakkies from c. 4 700 to 3 200 cal yr BP, although a marked arid episode is observed between c. 3 800 and 3 700 cal yr BP (Chase et al. 2015). The KB-1 pollen record exhibits corresponding moister conditions from c. 3 700 to 1 700 cal yr BP, with peaks in *Podocarpus* from c. 3 500 to 1 600 cal yr BP (Meadows et al. 2010). This record does however highlight opposing drier phases, especially after c. 3 600 cal yr BP as evidenced by the presence of Crassulaceae and Euphorbiaceae, and a notable reduction in Cyperaceae (Meadows et al. 2010). In this regard, Meadows et al. (2010) emphasises the fact that interpretations based on the KB-1 pollen record tend to be problematic as both drier and wetter indicator taxa are present together – it is suggested that this disparity is possibly due to the sampling resolution.

From c. 3 000 cal yr BP, increased $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ values in the Truitjies Kraal record point to warmer conditions as well as increased aridity (Meadows et al. 2010). Furthermore, slightly enriched $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ values are shown at De Rif from c. 2 300 to 950 cal yr BP indicating decreased moisture availability for this period (Quick et al. 2011). In contrast to this, wetter periods were evident at Pakhuis Pass at c. 2 800 and 2 000 cal yr BP, although drier conditions seemed to prevail amid these wet intervals (Scott & Woodborne 2007a, b). It is proposed that these wet phases are possibly due to the summer rainfall zone extending to its most southerly positions around c. 2 000 cal yr BP (Scott & Woodborne 2007a, b). A rapid intensification in summer rainfall is shown at Katbakkies at c. 2 700 cal yr BP, with a similarly moister environment indicated – this persisted until around c. 1 600 cal yr BP (Chase et al. 2015). Likewise, the KB-1 record implies wetter conditions c. 2 400 to 1 300 cal yr BP, with lower stable isotope values and increased proportions of Cyperaceae pollen (Meadows et al. 2010).

This wetter phase is followed by a shift towards a drier environment approaching the end of the record c. 600 cal yr BP (Meadows et al. 2010). At De Rif (DR-1) the prevalence of dry and/or warm indicator taxa, such as Aizoaceae, *Anthospermum*- and *Pentzia*- type similarly point to drier conditions from around c. 2000 cal yr BP to the top of the record at c. 700 cal yr BP (Quick et al. 2011).

At Pakhuis Pass, a noticeable decline in $\delta^{13}\text{C}$ values are observed from c. 2 000 to 1 200 cal yr BP (Scott & Woodborne 2007b). Together with the pollen records, Scott & Woodborne (2007b) interpret this to indicate a shift from wet summer-rain conditions to a drier environment - further substantiating the notion of drier climatic conditions as indicated above.

Moister conditions persisted at Pakhuis Pass from c. 1 000 to 400 cal yr BP (Scott & Woodborne 2007b) with evidence from Katbakkies similarly pointing to a return of a wetter environment during the last 1000 years of the record (Chase et al. 2015).

3.3.2. The Summer Rainfall Zone (SRZ)

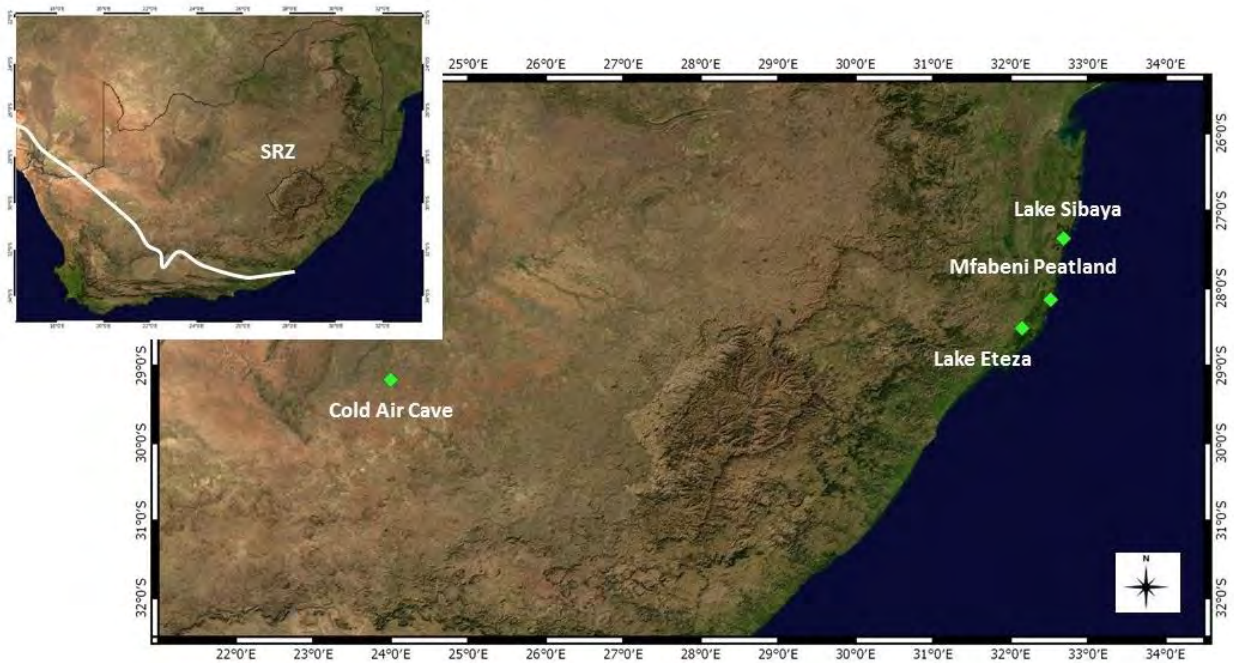


Figure 3.6. The locations of the sites from the summer rainfall zone discussed in the text

The East Coast

Recent records from the East Coast of South Africa, for example Lakes Sibaya and Eteza and the Mfabeni Peatland, provide a comprehensive view on the late Holocene in this region. The record from Lake Sibaya extends to c. 7 700 cal yr BP but a 5 000 year hiatus essentially shortens the record to mainly focus on the Iron Age from c. 1 500 cal yr BP (Neumann et al. 2008). A new high-resolution diatom record from Sibaya focussing on the last 1 800 year (Stager et al. 2013), further contributes to the late Holocene record. The Lake Eteza pollen record (Neumann et al. 2010) covers most of the Holocene with the record extending to c. 10 200 cal yr BP.

During the mid-Holocene, c. 6 800 to 3 600 cal yr BP, humid conditions and increased precipitation prevailed along the East Coast as evidenced by an increase in forest taxa in the Lake Eteza record

(Neumann et al. 2010). A sea level high stand is further inferred for this period with a concurrent peak in sea surface temperatures (Neumann et al. 2010).

Neumann et al. (2010) reports a decline in *Podocarpus* and other forest taxa after c. 3 600 cal yr BP, and an increase in grassland taxa such as Poaceae and *Phoenix* sp. Thus, indications are that a dry period persisted from c. 3 600 to 2 000 cal yr BP characterised by more open grasslands and a retraction of forest. Drier, cooler conditions are also observed at the Mfabeni peatland for the period c. 3 000 to 2 500 cal yr BP with a steady decline in *Podocarpus* evident in the pollen record (Finch & Hill 2008).

Moist coastal forest components return to the Eteza assemblage after c. 2 000 cal yr BP while a decrease is observed in the open grassland indicators Poaceae, *Phoenix* and Asteraceae pointing to a return to a more humid environment with increased moisture availability (Neumann et al. 2010). Interestingly, Stager et al. (2013) highlights considerably low values for the precipitation-evaporation index (P-E) for the period 1 800 to 1 790 cal yr BP and 1 755 cal yr BP, possibly suggesting very brief dry intervals punctuating this generally wetter period.

At the resumption of deposition in the Sibaya record, arboreal taxa, especially *Podocarpus*, is well represented from 1 500 to 1 000 cal yr BP indicating, as in the Eteza record, increased humidity but possibly cooler conditions (Neumann et al. 2008). Higher P-E values are also noted in the diatom record c. 1 600 to 1 555 cal yr BP, which could substantiate the suggested wetter environment (Stager et al. 2013).

Steady drying over this period is however evident in the records, with Lake Sibaya experiencing lower lake levels between c. 1 550 and 1 160 cal yr BP (Stager et al. 2013) and a marked decrease in forest components, and again specifically *Podocarpus*, are observed around c. 1 500 to 1 250 cal yr BP, with a concurrent expansion of grasslands (Neumann et al. 2008). A major increase in the P-E record, and lake levels, is indicated for the period c. 1 160 to 1 140 cal yr BP highlighted as the “most extreme rise in the record” suggestive of a major wet event (Stager et al. 2013, page 114) during this generally drier period.

From c. 800 cal yr BP, savanna conditions developed along the East Coast as indicated by a sudden increase in *Spirostachys africana* pollen in the Lake Eteza record (Neumann et al. 2010). Peaks in Cyperaceae and fern spores possibly suggest open water, swampy conditions likely associated with lower lake levels Neumann et al. (2010). Significantly low P-E values for the period c. 770 to 750 cal yr BP are indicated by the diatom record, further supporting the notion of lower lake levels (Stager et al. 2013).

Regional moister conditions are suggested between c. 850 and 350 cal yr BP from the Sibaya pollen record, although notably low P-E values are recorded from c. 770 to 750 cal yr BP at Sibaya (Stager et al. 2013). A retraction of arboreal elements, are evident from c. 700 cal yr BP (Neumann et al. 2008) with drier conditions further suggested from the Lake Eteza record c. 700 years ago (Neumann et al. 2010) as well as the Lake Sibaya conductivity series and gray scale record, with the P-E record further pointing to significantly decreased values c. 560 cal yr BP (Stager et al. 2013). This might be the expression of the cooler and/or drier conditions during the Little Ice Age, which persisted between 650 and 150 cal yr BP (Stager et al. 2013). A concurrent prompt increase in sedimentation rates c. 700 cal yr BP possibly marks the onset of agricultural activities by Iron Age settlers, which possibly further contributed to a decrease in forest taxa (Neumann et al. 2008; Neumann et al. 2010). A distinctly wetter period prevailed in the area c. 480 to 410 cal yr BP followed by notably dry conditions extending to 190 cal yr BP (Stager et al. 2013).

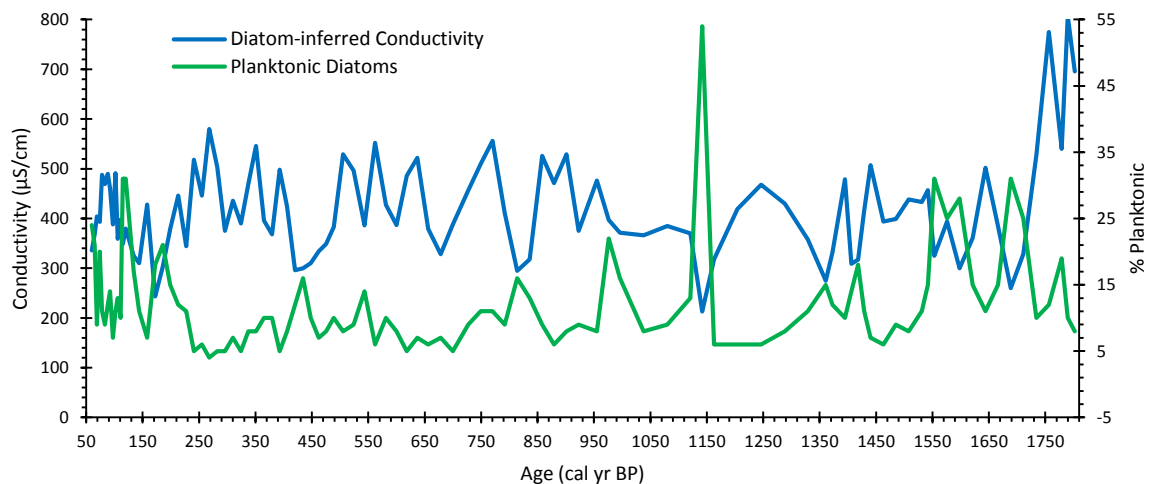


Figure 3.7. c. 1800 year diatom record from Lake Sibaya (after Stager et al. 2013)

Stager et al. (2013) suggests further that in this region, the Medieval Climate Anomaly was largely wetter than average while the LIA was drier. Significant wet and dry periods did however occur intermittently throughout these two events suggesting highly variable hydroclimatic conditions for the period c. 950 to 150 cal yr BP (Stager et al. 2013).

Neumann et al. (2010) highlight dramatic changes in vegetation from c. 700 cal yr BP, most notably an increase in *Pinus* pollen, at the expense of other forest elements. This would mark the arrival of European settlers in the region during the 19th Century (Neumann et al. 2008; Neumann et al. 2010). *Stoebe*-type and *Chenopodiaceae* pollen are also abundant in the records for this period, suggesting

disturbances in the environment, most probably due to anthropogenic activities (Neumann et al. 2008; Neumann et al. 2010). The presence of neophyte pollen, such as *Casuarina* and *Ambrosia*-type, further points to the influence of Colonial settlers on the surrounding environment during the last c. 200 years (Neumann et al. 2010).

Cold Air Cave, Makapansgat Valley

Towards the interior of the country, the stalagmite based isotope record from Cold Air Cave is put forward as being the most comprehensive record to incorporate Late Holocene climate change in South Africa, since the Talma and Vogel (1992) record from the Cango Cave stalagmite (Tyson 1999a, b; Holmgren et al. 2003).

Although a variety of records, representative of the interior of the SRZ are available (e.g. Tswaing crater), Cold Air Cave was chosen due to its comparability with the Cango Cave records (Talma & Vogel 1992) – which forms an integral part of the palaeorecords from the YRZ.

The middle Holocene in the interior of the country, is indicated to have been wetter, humid and relatively warm as evidenced by positive $\delta^{18}\text{O}$ and lower $\delta^{13}\text{C}$ values from the isotope record (Lee-Thorp et al. 2001). The lower $\delta^{13}\text{C}$ values further suggest increased vegetation cover in the form of trees and/or shrubs, while C4 grasses are scarce (Lee-Thorp et al. 2001). A trend towards cooler conditions are however evident as $\delta^{18}\text{O}$ values show a steady decline from around c. 6 000 to 2 500 BP (Holmgren et al. 2003).

Variability in temperature and moisture availability is shown for the period c. 4 300 to 3 200 BP, after which a notable change in both climate and vegetation is observed in the $\delta^{13}\text{C}$ record as well as the grey index series (Lee-Thorp et al. 2001). From this point, c. 3 200 BP, a trend towards increasing aridity and cooling is evident until c. 2 100 BP, with minimum $\delta^{18}\text{O}$ values, and thus drier conditions, expressed at c. 3 100 and 2 600 BP. Holmgren et al. (1999) further emphasises notable cool periods in the record between c. 2 950 and 2 750 BP and again between c. 2 150 and 1 950 BP. Maximum enrichment of $\delta^{18}\text{O}$ is however observed around c. 2 300 BP which is linked to progressively warmer conditions (Holmgren et al. 1999; Tyson 1999a). The $\delta^{13}\text{C}$ record correspondingly increases to peak at around c. 2 100 BP, emphasising the ample presence of C4 grasses in the environment (Lee-Thorp et al. 2001).

Between 2 000 and 1 000 BP alternating cooler and warmer periods occurred. Significant warm spells are evident at c. 1 950 to 1 850 BP and c. 1 650 to 1 550 BP while discrete cold events are shown to have prevailed from c. 1 510 to 1 430 BP and c. 1 150 to 1 050 BP (Holmgren et al. 1999; Tyson 1999a, b). The $\delta^{18}\text{O}$ record further points to primarily wetter and warmer conditions from c. 1 950 to 1 650

BP with this period identified as being the most persistent and constantly warm period in the Cold Air Cave record (Holmgren et al. 1999; Tyson 1999a, b).

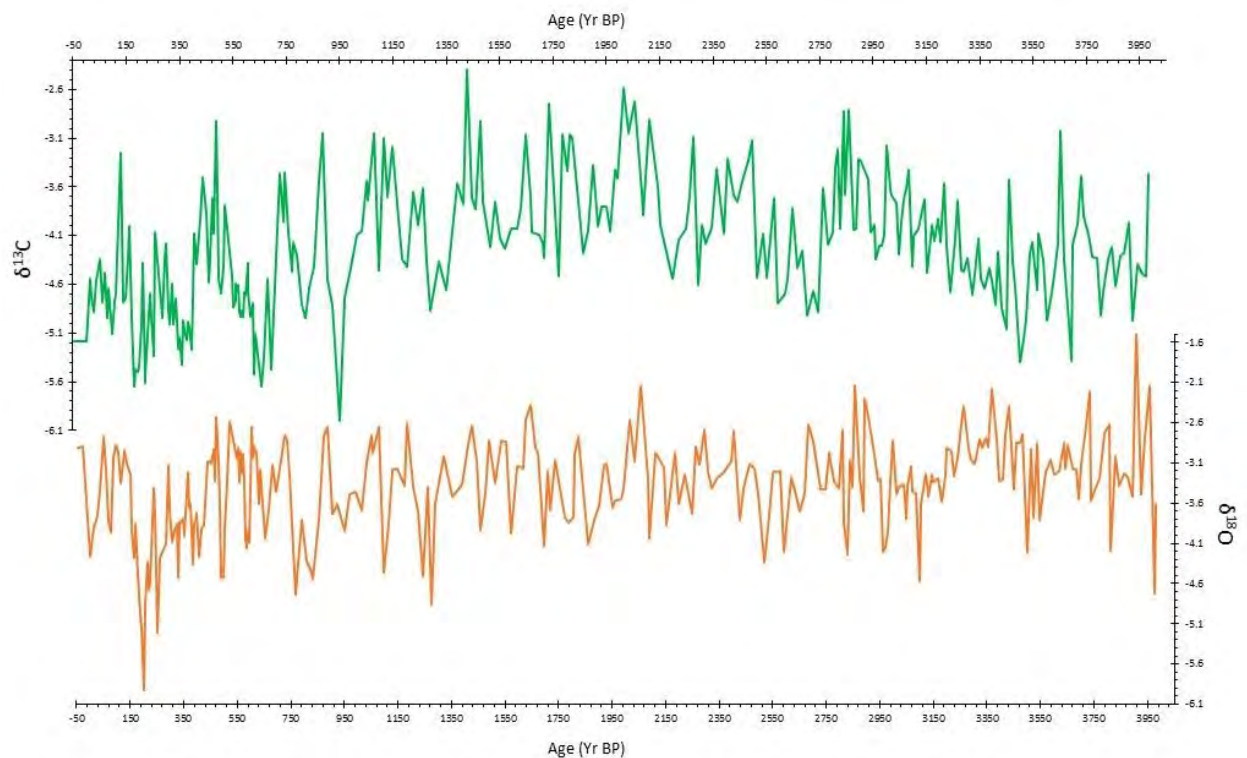


Figure 3.8. The late Holocene $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ isotope records from Cold Air Cave (after Holmgren et al. 2003)

The Medieval Climate Anomaly (MCA) in the interior of the country prevailed from c. 1 050 to 650 BP (Holmgren et al. 1999; Tyson 1999a, b; Tyson et al. 2000) and is (described) as a generally warmer, highly variable period with intermittent brief cool events (Holmgren et al. 1999; Tyson 1999a, b). Variations in the grey level record are indicated to be above average with the $\delta^{18}\text{O}$ record similarly indicating warmer and possibly wetter conditions for this period (Holmgren et al. 1999; Tyson 1999a, b). Substantial variability is observed in both the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ records with the most rapid fluctuations evident between 770 and 730 BP (Holmgren et al. 1999). This corresponds to the timing of the transition from the MCA to the Little Ice Age (LIA) (Holmgren et al. 1999; Tyson 1999a, b).

The major event in the CAC record is the cool, dry period between c. 650 and 150 BP, which Holmgren et al. (1999) identifies as the expression of the LIA in South Africa. A brief sporadic warmer phase around c. 450 to 275 BP essentially divides the LIA in to two phases. From here, temperatures decrease to their lowest levels at c. 250 BP where the most distinct negative excursion of $\delta^{18}\text{O}$ values are seen (Holmgren et al. 1999; Holmgren et al. 2003). By c. 140 BP, the LIA came to an abrupt end,

marked by rapid enrichment in the $\delta^{18}\text{O}$ series (Tyson 1999a, b; Holmgren et al. 1999; Tyson et al. 2000).

3.3.3. The Year Round Rainfall Zone (YRZ)

The YRZ provides an interesting study area as through time the region would have been dominated by alternating summer and winter rainfall regimes while also exhibiting unique characteristics – in relation to the WRZ and SRZ – when the year round rainfall system became established.

Various studies focusing on the YRZ are available, though only a hand full focus on the late Holocene or include a complete late Holocene section. With regard to these records Chase and Meadows (2007) emphasise some important issues concerning inferences made from this evidence; the most significant of these being the fact that the southern coast falls within the YRZ while also being part of the Fynbos Biome which is partially situated within the WRZ. Evidence from this region thus tend to be interpreted from a WRZ perspective albeit that there is substantial difference between the climatic conditions in these two zones.



Figure 3.9. The locations of the sites from the year round rainfall zone discussed in the text

The Southern Cape Coast

Norga Peat, George

A c. 4 500 year long pollen record originated from the Norga peat located near the town of George. Even though this record is based on only three radiocarbon dates and results were never officially published, this work from Scholtz (1986) forms an integral part of the Southern coast palaeoenvironmental record – mostly due to the limited amount of information available for this region.

Three distinct climatic phases were identified from the record. The first covers the period c. 4 400 to 2 600 cal yr BP and is marked by ample summer rainfall and a moderate climate conducive to the expansion of Afrotemperate forest elements (Scholtz 1986). This is followed by an observed decline in forest elements, i.e. *Podocarpus*, *Olea* and *Ilex*, from c. 2 600 to 1 300 cal yr BP. Scholtz (1986) ascribes this change to a decrease in summer rainfall resulting in a drier and cooler environment less suitable for forest development. Included in this time frame, Scholtz (1986) presents the period c. 2 500 to 1 700 cal yr BP to represent “the most adverse conditions experienced in the southern Cape during the last ca. 4 000 years” (page 11). Scholtz's (1986) ‘last phase’ of forest expansion seems to have started between c. 1 300 and 850 cal yr BP pointing to the development of a more mesic, and possibly warmer, climate. This also marks the last climatic phase, extending to the present day, during which contemporary climatic conditions became established in the region (Scholtz 1986).

Nelson Bay Cave

Cohen & Tyson (1995) presented a Holocene sea surface temperature (SST) record from the Agulhas Bank. It was obtained by evaluating the oxygen isotope ($\delta^{18}\text{O}$) composition in marine mollusc shells from deposits from Nelson Bay Cave.

The record shows that, for the period c. 6 800 to 5 400 cal yr BP, both summer and winter SST's were higher than present along the Agulhas Bank (Cohen & Tyson 1995). Reasonably lower summer maximum SST's are indicated from c. 4 800 cal yr BP, with a decrease in maximum temperatures of around 1°C, in relation to today, experienced between c. 3 500 and 2 500 cal yr BP (Cohen & Tyson 1995). Lower summer and winter SST's are again shown at c. 650 cal yr BP, possibly an indication of the onset of the Little Ice Age, with a marked decline of between 1 and 2 °C noted at c. 600 cal yr BP (Cohen & Tyson 1995).

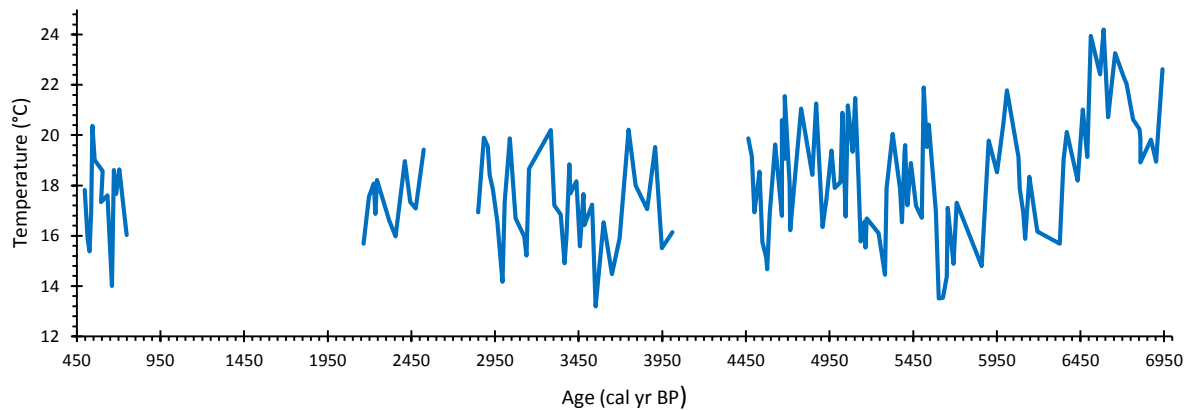


Figure 3.10. Holocene sea surface temperature record from the south coast (after Cohen & Tyson 1995)

Cohen & Tyson (1995) further used this data to test the regional atmosphere-ocean model first proposed by Tyson (1986). According to the model, lower SST's along the Agulhas Bank are related to increased upwelling along this part of the coast. In association with colder SST's in the southern Benguela, these conditions are probably related to increases in rainfall in the SRZ and drier conditions along the west coast (Cohen & Tyson 1995). It is further proposed that lower SST's along the south coast are driven by the increased influence of the easterlies and a concurrent reduction in westerly flow (Cohen & Tyson 1995; Chase et al. 2013).

Vankervelsvlei

This unique wetland (see Quick, 2013 for a detailed description) has unfortunately not delivered a full Holocene record as of yet, with the pollen record from Irving (1998) only extending to c. 3 300 cal yr BP and the Quick (2013) record only covering the early Holocene. Quick (2013) further highlights some issues regarding the Irving (1998) record pointing to “poor chronological control” and problems with taxonomic identification leading to unreliable results. Irving (1998) does however, as with Martin (1968) and Scholtz (1986), identify the period c. 4 400 to c. 3 200 cal yr BP as being favourable for the development of forests. According to Quick (2013) forest elements are present in the Irving (1998) record but these rather point to coastal thicket or scrub forest components instead of Afrotropical elements.

The Wilderness Embayment

Until recently, the only evidence from the Wilderness Embayment was that of Martin (1956, 1959, 1960, 1962, 1968) with a strong focus on Groenvlei. A new late Holocene diatom record from Eilandvlei (Kirsten 2014) is however now available.

The Groenvlei pollen record extends to c. 8 800 cal yr BP spanning the period from the Holocene Altithermal to the present and includes the era of European occupation, with the Eilandvlei diatom record covering the last 4000 years.

A strong marine influence is observed in Groenvlei from c. 7 800 to c. 3 200 cal yr BP indicated by increased levels of *Chenopodiaceae* and *Restionaceae* pollen in the sequence (Martin 1968). Martin (1968) suggests that a rise in the ground-water table, due to sea-level rise, might be responsible for the observed high *Restionaceae* levels. This marine effect is also present in Eilandvlei until c. 3 750 cal yr BP with Kirsten (2014) suggesting that the marine intrusion was at a maximum around c. 3 900 cal yr BP.

A marginally drier period in the region is proposed from c. 4 000 to 3 850 cal yr BP, followed by a substantially wetter phase between c. 3 750 and 3 650 cal yr BP (Kirsten 2014). A momentary dry event is indicated for the 100 years between c. 3 650 and 3 550 cal yr BP resulting in lower lake levels at Eilandvlei before a recovery at around c. 3 500 cal yr BP (Kirsten 2014). Kirsten (2014) further defines the period between c. 3 750 and 3 500 cal yr BP as “climatically variable”. This variable period corresponds with widespread Aeolian deposition and the onset of late Holocene dune sedimentation in the region between c. 3 700 and 2 400 cal yr BP (Bateman et al. 2011). During this period and extending to c. 1 900 cal yr BP, Martin (1968) also puts forward geomorphological and sedimentological evidence indicating movement of the surrounding sand dunes. The most probable explanation for this being receding sea levels after the high stand observed during the mid-Holocene (Scholtz 1986; Bateman et al. 2011). In this regard, Martin (1968) cautions interpretations based on the expansion and/or contraction of forests during this period, as observed in the pollen record, as extensive dune activity might have been responsible for deforestation in the area – variations that could similarly be attributed to changes in climate.

Two climate scenarios are consequently presented for this period: conditions were reasonably dry and/or warmer creating an environment not suitable for forest expansion; alternatively it may have been wetter but windier, with moving dune sands the restricting factor with regard to forest spread, and not climate (Martin 1968). Drier conditions are however shown to prevail until c. 2 100 cal yr BP, with the lowest lake levels in the Eilandvlei sequence recorded at around c. 2 375 cal yr BP (Kirsten 2014).

A transitional phase is identified in the Groenvlei pollen record from c. 1 900 to c. 1 800 cal yr BP, characterised by a decline in herbaceous taxa, i.e. Ericaceae and *Anthospermum*, while an increase in arboreal pollen is noted. *Podocarpus* does however display minimum values here, followed by a prompt increase from c. 1 800 cal yr BP – indicating the onset of the so-called last phase of forest development (Martin 1968). This possibly points to a return to more mesic conditions with environmental conditions more suitable to the expansion of the forests. This transitional phase is similarly evident in the Eilandvlei record between c. 2 100 and 2 000 cal yr BP with Eilandvlei changing from a lagoon to a coastal lake (Kirsten 2014). This period further displays rapid changes in climatic conditions moving from significantly increased aridity to substantially wetter conditions within a short space of time (Kirsten 2014). Increased moisture availability is further indicated at c. 2 025 cal yr BP accompanied by evidence relating to a significant flooding event in the Eilandvlei catchment at the same time (Kirsten 2014).

Accordingly, indications are that a wetter environment persisted until about c. 1 700 cal yr BP after which a cooler and/or windier, and possibly more arid, interval prevailed from c. 1 750 to 1 350 cal yr BP (Kirsten 2014). These conditions are further substantiated by dune movement around Sedgfield, also from around c. 1 750 to 1 300 cal yr BP (Bateman et al. 2011).

Drier conditions, supplemented by restricted moisture availability, are again evident from c. 1 200 cal yr BP in the diatom record, with increased aridity between c. 1 000 and 825 cal yr BP (Kirsten 2014). A slight regression in arboreal taxa is shown at Groenvlei around c. 850 cal yr BP, as well as significantly increased levels of Restionaceae and Asteraceae (undefined) (Martin 1968). This is suggested to be indicative of the fact that the onset of forest fragmentation predated European settlement, probably in response to a drier climate (Martin 1968). Kirsten (2014) further posits that this dry interval corresponds to the global timing of the Medieval Climate Anomaly.

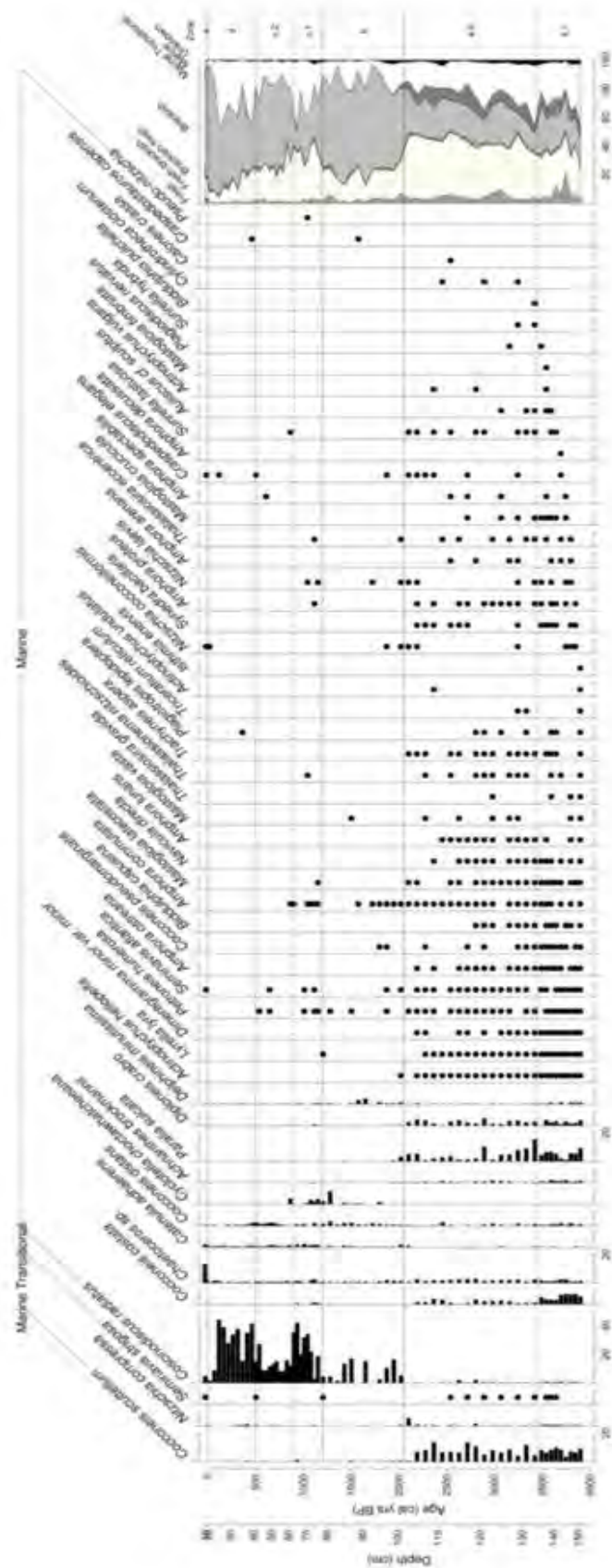


Figure 3.11.2. ~4000 year diatom record from Eilandvlei continued (Kirsten 2014)

Wetter conditions returned to the region around c. 800 cal yr BP and was sustained until c. 550 cal yr BP (Kirsten 2014). Kirsten (2014) posits that recurrent winter rainfall in conjunction with prevailing summer rainfall in the region, are responsible for this wet interval. A complementary, shorter, diatom record from nearby Swartvlei likewise points to largely wetter conditions c. 700 to 500 cal yr BP (Kirsten 2014). The Eilandvlei record further points to a possible increased marine influence in the lake c. 600 to 500 cal yr BP (Kirsten 2014). This wet episode is most probably the expression of the onset of the Little Ice Age along the south coast (Kirsten 2014).

Reinwarth et al. (2013) note the increasing concentration of fine minerogenic sediment components in Eilandvlei from around c. 500 cal yr BP and suggests that these changes in sediment dynamics were most possibly driven by changes in climate as well as fluctuating sea levels. A resurgence in dune activity at Sedgefield is observed from c. 600 to 430 cal yr BP (Bateman et al. 2011) suggesting that conditions along the south coast seem to be highly dynamic and variable at the onset and during the first phase of the LIA.

Lower lake levels are registered at both Eilandvlei and Swartvlei between c. 500 and 300 cal yr BP with the availability of freshwater seeming to be the major influencing factor (Kirsten 2014). In regard to this, Kirsten (2014) suggests that this period agrees with the warm episode punctuating the two cooler phases of the LIA. A second dry episode is observed at c. 220 to 190 cal yr BP after which, a marginal increase in freshwater input is noted until c. 140 cal yr BP (Kirsten 2014).

The Southern Cape Interior

Cango Cave, Oudtshoorn

Before the record from Cold Air Cave was published, the c. 6 000 year Cango cave ^{18}O isotope record was the longest Holocene temperature record available for southern Africa (Tyson & Lindesay 1992; Tyson & Preston-Whyte 2000).

From the onset of the Holocene portion of the Cango Cave sequence, c. 5 900 cal yr BP, the temperature record, derived from the $\delta^{18}\text{O}$ series, remains relatively stable, only registering fluctuations between +1 and -2 °C (Talma & Vogel 1992). Greater variability is observed in the $\delta^{13}\text{C}$ record, which exhibits a steady increasing trend from c. 5 900 cal yr BP, reaching a maximum at around c. 2 000 cal yr BP (Talma & Vogel 1992).

Talma & Vogel (1992) infers this $\delta^{13}\text{C}$ peak to point to an increase in C4 vegetation in the region which according to Vogel (1978; as cited by Talma & Vogel 1992), is the result of increased average

temperatures during the usually wet, growing season. Scott (1993) similarly proposes that this peak points to the replacement of C3 plants by C4 vegetation related with an increase in summer rainfall. It is further posited that this could be the expression of the southernmost movement of the SRZ (Scott 1993). In contrast to this, Talma & Vogel (1992) hypothesises that the disparity between the constant $\delta^{18}\text{O}$ and more variable $\delta^{13}\text{C}$ records is the result of a change towards increased winter rainfall seasonality, extending to around c. 2 000 cal yr BP (Talma & Vogel 1992).

Even though the temperature record is deemed to reflect relatively constant conditions, variations are present which tend to correspond with globally recognised climatic events. Generally lower temperatures are indicated between c. 5 900 and 2 500 cal yr BP with two major cool intervals highlighted from c. 5 400 to 4 800 cal yr BP, and from c. 3 350 to 2 600 cal yr BP (Talma & Vogel 1992). Tyson and Preston-Whyte (2000) relates the second cool event with the neoglaciation, a period of global glacial expansion. Following this period of neoglaciation cooling, a steady increase in temperature is observed culminating with the maximum value within the record at c. 2 090 cal yr BP (Talma & Vogel 1992).

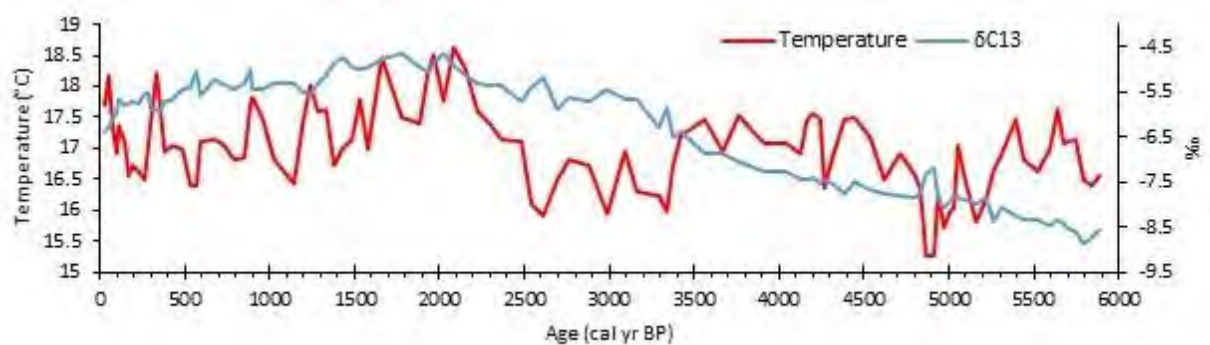


Figure 3.12. The Congo Cave temperature and $\delta^{13}\text{C}$ records (after Talma & Vogel 1992)

The 200 years around this peak in temperature is indicated to be warm and wet, followed by rapid cooling from c. 1 850 to 1 700 cal yr BP (Talma & Vogel 1992; Tyson 1999b). Evidence again points to a warmer environment from c. 1 700 to 1 350 cal yr BP, followed by colder conditions until around c. 1 050 cal yr BP (Talma & Vogel 1992; Tyson & Lindesay 1992). A peak in temperature is noted at the end of this cooler phase, marking the onset of the Medieval Climate Anomaly in the region (Talma & Vogel 1992; Tyson 1999b). This period is defined as being warm and variable, extending to around c. 650 cal yr BP where some of the lowest temperatures in the record are recorded (Talma & Vogel 1992; Tyson & Lindesay 1992). It is thus inferred that the Little Ice Age started at this point, c. 650 cal yr BP, which Tyson & Lindesay (1992) identified as the coldest phase of the LIA. Cooler conditions persisted until a prominent warm event is noted around c. 340 cal yr BP (Talma & Vogel 1992; Tyson & Lindesay 1992).

A return to a colder environment is then noted, extending to c. 250 cal yr BP, followed by a prompt increase in temperature until the end of the record (Talma & Vogel 1992; Tyson 1999b).

Seweweekspoort

New Holocene isotope records from the southern cape interior has recently been made available. These records are derived from hyrax middens from Seweweekspoort, and together cover the mid-to late Holocene from c. 77 00 to 80 cal yr BP (Chase et al. 2013).

Distinct climatic variability is displayed in the record with alternating wetter and drier periods most notable in the $\delta^{15}\text{N}$ series (Chase et al. 2013). Consequently, increased humidity are inferred for the periods c. 7 300 to 6 600 cal yr BP, c. 6 200 to 57 00 cal yr BP and c. 5 200 to 4 500 cal yr BP (Chase et al. 2013). Phases of increased aridity are further evident from c. 7 800 to 7 300 cal yr BP, c. 6 800 to 5 500 cal yr BP and c. 4 500 to 3 400 cal yr (Chase et al. 2013)

In comparing their results with the sea surface temperature record from Cohen & Tyson (1995), Chase et al. (2013) found that drier, warmer phases relate to periods of lower SST's along the south coast. As Cohen & Tyson (1995) proposed that these lower SST's occur when the influence of the easterlies increase, westerlies decrease and upwelling along the coast is stronger, Chase et al. (2013) posits that these arid episodes at Seweweekspoort are a result of the reduced influence of the westerlies and their associated moisture-bearing systems.

3.3.4. Key Trends in the South African Palaeoenvironmental Records

Following the discussion on the palaeoenvironments of the three rainfall zones of South Africa, Tables 3.1-1 to 3.1-3 provides a summary of the key trends during the late Holocene from these regions.

Table 3.1-1 Summary of the key trends from the three rainfall zones of South Africa between c. 4000 and 2000 cal yr BP (PV = Princessvlei; RV = Rietvlei; CAC = Cold Air Cave; WE = Wilderness Embayment; SWP = Seweweekspoort)

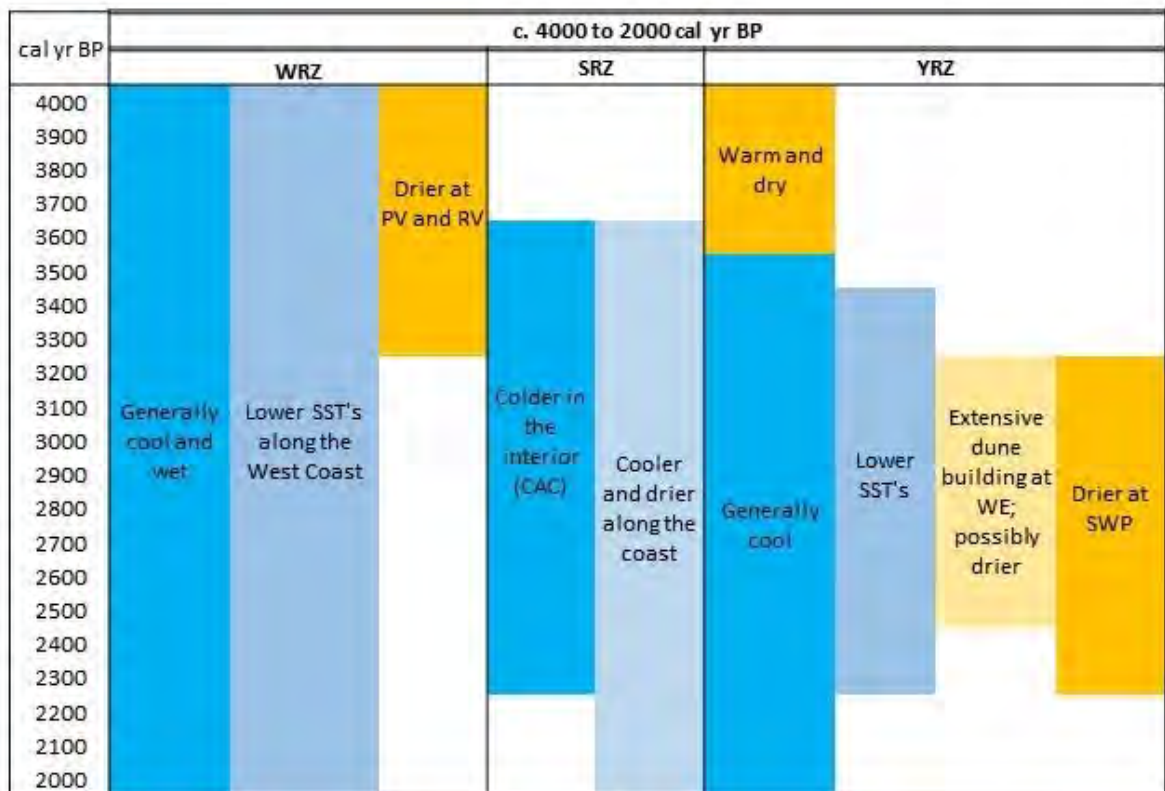


Table 3.1-2 Summary of the key trends from the three rainfall zones of South Africa during the Medieval Climate Anomaly (PP = Pakhuis Pass; KB = Katbakkies; PV = Princessvlei; CAC = Cold Air Cave; WE = Wilderness Embayment; CC = Congo Cave; SWP = Seweweekspoort)

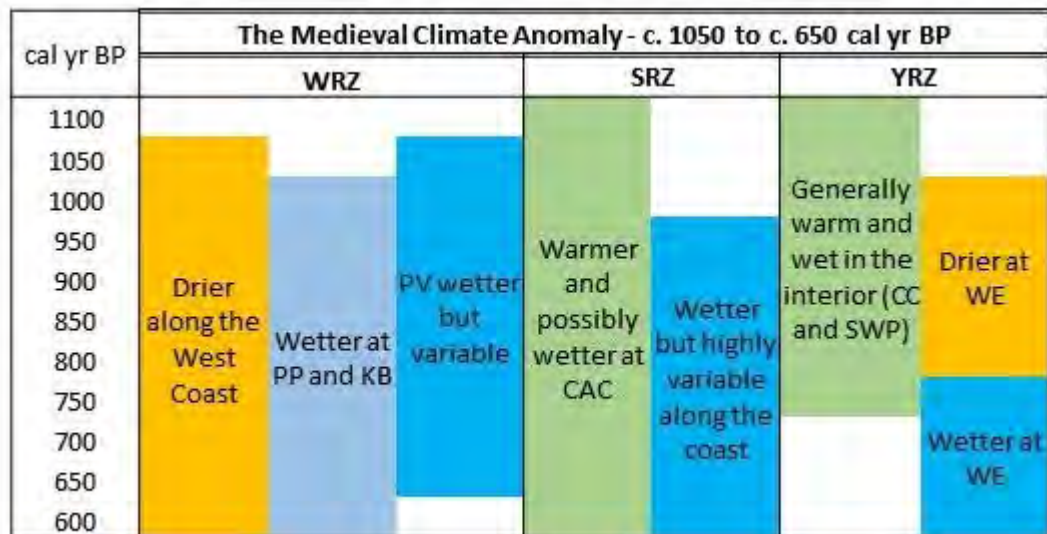
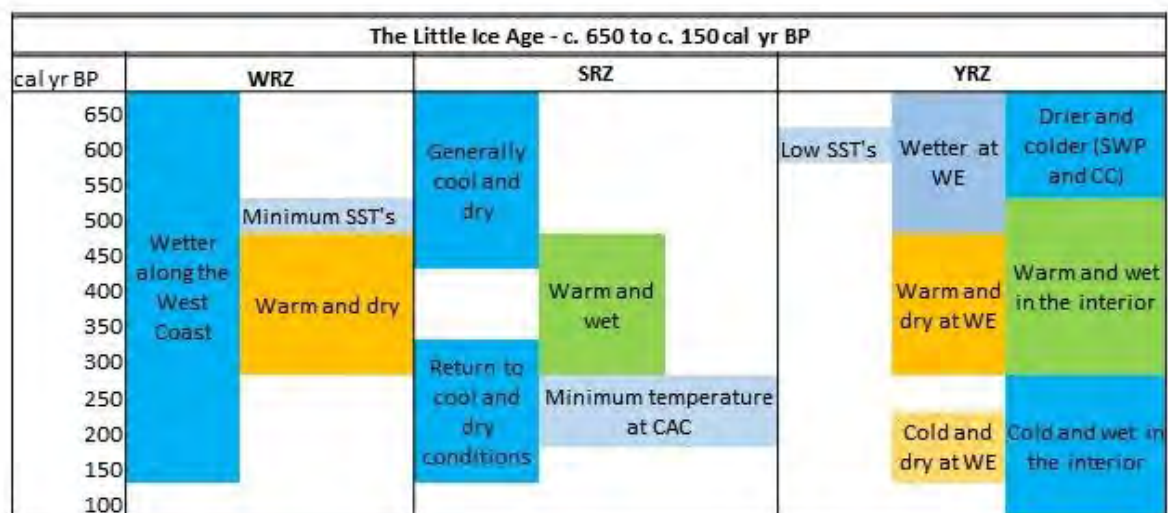


Table 3.1-3 Summary of the key trends from the three rainfall zones of South Africa during the Little Ice Age (CAC = Cold Air Cave; WE = Wilderness Embayment; CC = Congo Cave; SWP = Seweweekspoort)



3.4. Conclusion

This chapter illustrated the regional climatic and environmental differences in South Africa during the Late Holocene. Furthermore, this chapter provides a theoretical basis upon which the Eilandvlei record can be investigated (Chapter 6).

4. Methodology

4.1. Introduction

This chapter outlines the methodological approach taken to obtain proxy source material – with a specific focus on lake sediments.

4.2. Theoretical Considerations

4.2.1. Lake Sediment Proxy Sources and Deposition

Lake sediments provide an invaluable archive of palaeoecological information, essentially acting as a sediment trap, preserving proxy material contained therein.

The influx of pollen grains into lake sediments are facilitated by various mechanisms (Figure 4.1) including sediment and water transport through surface runoff and erosion (Cw). Local aquatic plants within the lake (Cl), further contribute to pollen accumulation, while the majority of the pollen grains enter the lake through atmospheric transport, or pollen rain (Ac).

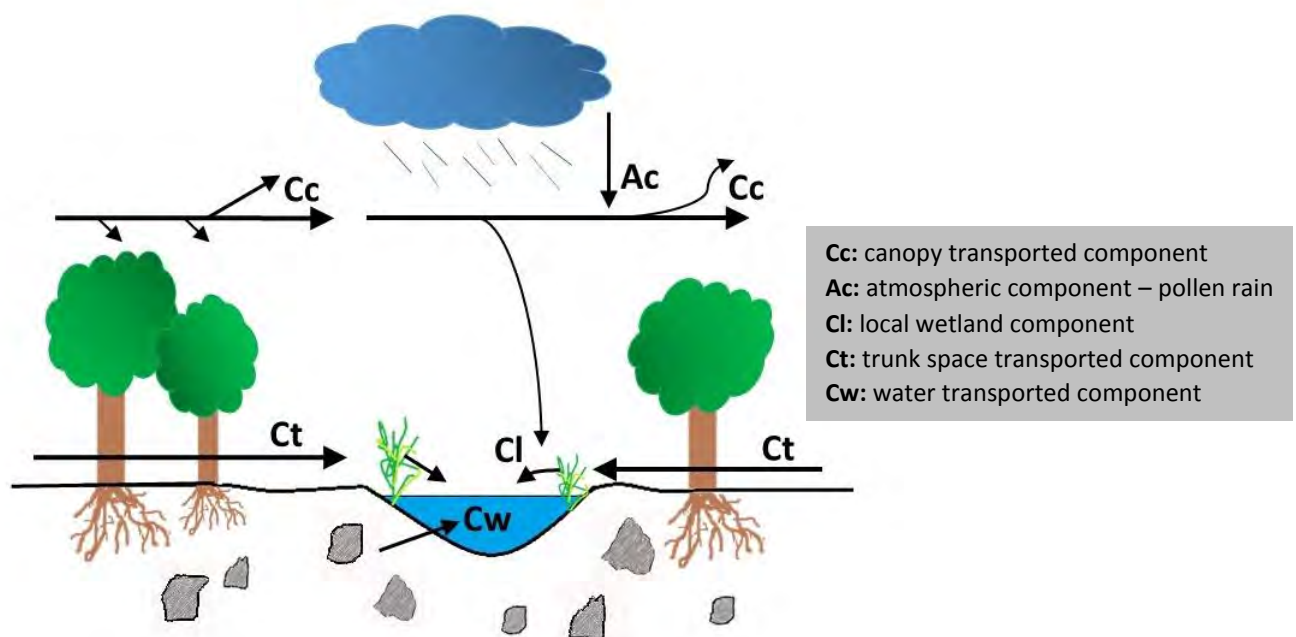


Figure 4.1 Pollen influx into lake sediments (adapted from Moore et al. 1991)

Charcoal deposition follows much the same mechanisms as pollen, with wind, or air convection, as the primary source of dispersion (Figure 4.2) (Conedera et al. 2009). This does however have the effect of incorporating charcoal particles from various ranges outside the catchment.

Other factors influencing the deposition, and thus accumulation, of charcoal within sediments include the geomorphology of the surrounding area, weather conditions and the type of plant species involved (Conedera et al. 2009). With sediment mixing and re-suspension further affecting charcoal deposition.

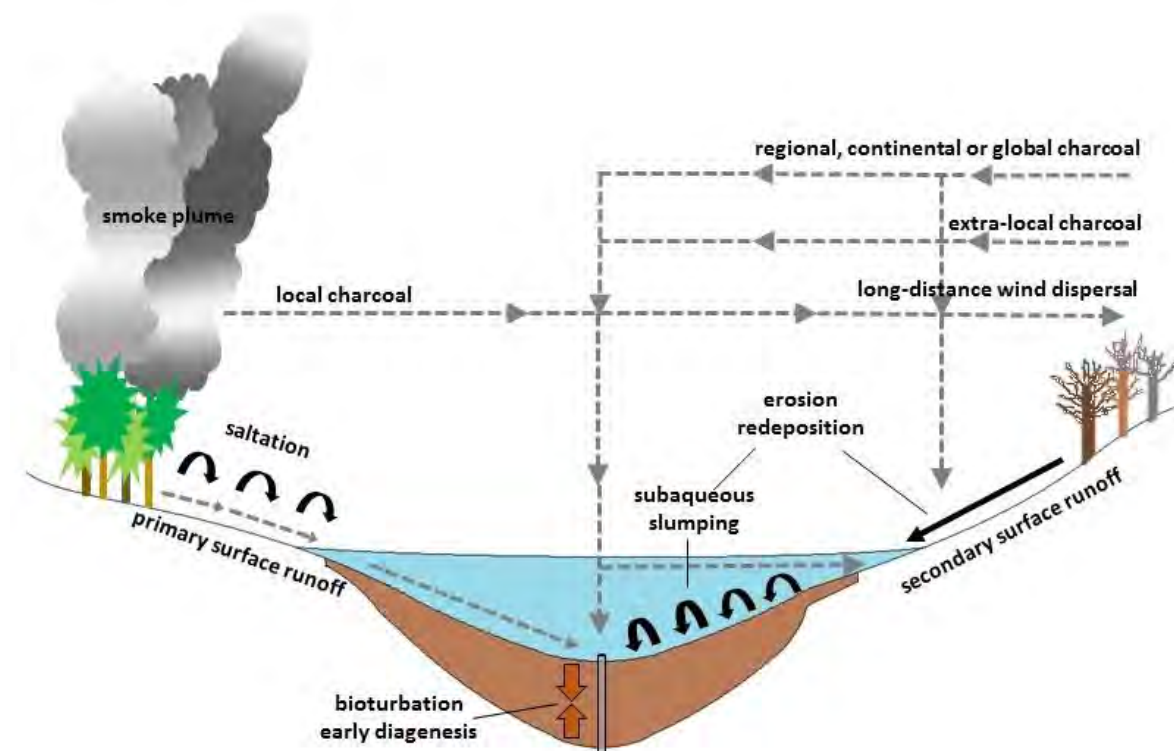


Figure 4.2 Sources of Charcoal Particles (redrawn from Conedera et al. 2009)

4.2.2. Pollen Analysis

Pollen analysis, or palynology, is one of the most widely applied methods for the investigation of Quaternary environments. Through the analysis of fossil pollen and spores questions regarding major vegetation change and ecosystem dynamics can be explored.

Pollen grains lend themselves perfectly to analysis over extended timescales, mainly due to the chemical composition which allows the preservation of the grain over millennia. The exine, or outer layer of the grain, contains the highly resistant material sporopollenin, protecting the pollen grain from microbial decay (Faegri & Iversen 1989; Traverse 2008). This outer layer is also responsible for the differing shapes and apertures (i.e. colpi and pores) making taxonomic identification of pollen grains possible (Moore et al. 1991).

This does not however mean that the method is without fault as there are inherent limitations when conducting pollen analysis. One of the most significant limitations to this method is the fact that identification of fossil pollen can typically only be made down to family level (Meadows & Sugden 1991), with possible damage to the grain, due to environmental conditions or during processing, further complicating the identification process. This particularly afflicts pollen analysis in southern Africa. Another important factor is the differing pollen production levels of individual taxa as well as the varying modes of dispersal of pollen grains (Figure 4.1) (Lowe & Walker 1997). Taxa with high pollen productivity as well as those dependent on wind dispersal (e.g. *Podocarpus*), tend to be over represented in the pollen record (Jackson & Williams 2004) which has to be taken into account, specifically when interpreting data. The depositional environment and accumulation of sediment, and thus pollen, needs to be considered as well when making inferences from pollen evidence with the proximity of vegetation to the site and the prevailing wind direction further to be taken into consideration (Moore et al. 1991).

4.2.3. Charcoal Analysis

The study of fossil charcoal particles in sediments provides information with regard to the incidence of past fires and fire regimes in the region from which the sediment is collected (Tinner & Hu 2003; Conedera et al. 2009). Together with pollen records, charcoal assemblages can further give insight into the fire-vegetation dynamics of the surrounding environment.

Several factors have to be taken into account when considering charcoal particles as a representation of past fire events within the local catchment. One of the most important of these is the uncertain source of microcharcoal particles. Charcoal calibration studies (e.g. MacDonald et al. 1991; Tinner et al. 1998) have shown that microcharcoal particles (10-200 μm) can originate from 20 to 100 km from the eventual deposition site thus presenting more of a regional fire signal. Macrocharcoal particles (100-200 μm), on the other hand, are typically not dispersed further than a few hundred meters from the fire providing evidence of local fire events (Tinner et al. 1998; Conedera et al. 2009). Further, pollen preparation procedures tend to breakdown charcoal particles leading to the over-representation of microcharcoal particles in the record.

Despite these aspects complicating the interpretation and reconstruction of past fire records, fossil charcoal can still be considered a valuable proxy, especially as supplement to pollen assemblages.

4.2.4. Radiocarbon Dating

Decades after W.F. Libby discovered radiocarbon dating, it is still one of the most commonly applied and reliable dating methods, especially relating to the Late Pleistocene and Holocene (Walker 2005; Hua 2009). This method utilizes the remaining amount of radiocarbon, or ^{14}C , present in all organisms, in the fossil sample. This is then compared to contemporary levels of ^{14}C in standard materials, from which an age for the fossil material can be established (Walker 2005). Due to the rate of decay, and thus the half-life of the ^{14}C atom of 5 730 years, this dating method has an age limit of $\sim 45\,000$ years, or eight half-lives (Walker 2005).

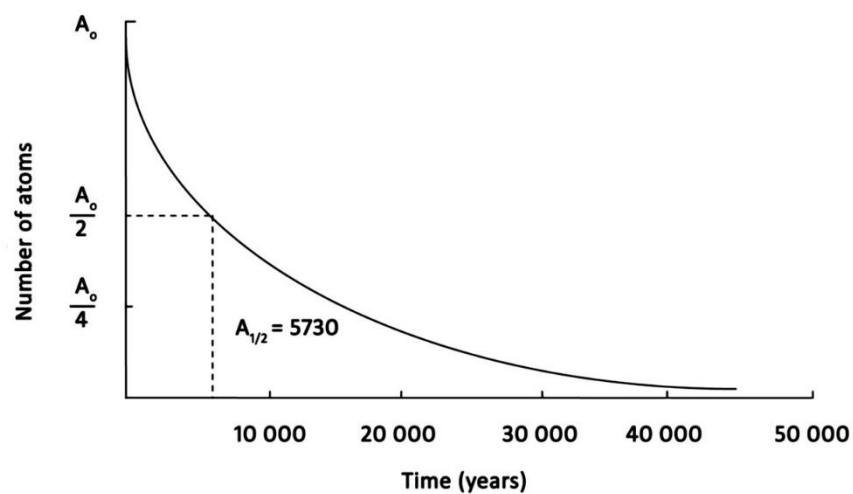


Figure 4.3 Curve demonstrating the rate of radiocarbon decay (from Walker 2005)

4.3. Field Sampling

4.3.1. Study Site

Forming part of the Wilderness Lakes complex, Eilandvlei comprises a surface area of 143 ha with a maximum depth of 6.5m (Allanson & Whitfield 1983). It serves to connect the Touw River system to the sea, through the Serpentine channel, although the lake experiences only intermittent, weak tidal exchanges when the mouth of the Touw River is open (Martin 1956; Allanson & Whitfield 1983; Russell 1999). Eilandvlei is additionally connected to Langvlei with freshwater input from the Touw and Duiwe Rivers as well as through underground seepage - variations in lake levels are thus attributed to rainfall and the extent of mouth closure (Allanson & Whitfield 1983; Russel et al. 2012). It has been found that the mouth of the Touw Estuary is open 30% of the time resulting in the variation

of salinity within the lakes (Russell et al. 2012). Lake salinity increases the further the lakes are situated from the sea: for example, Eilandvlei, being the closest to the ocean, has an average salinity of between 6 – 10 g/kg, Langvlei 10 – 13 g/kg, and Rondevlei 12 – 16 g/kg (Allanson & Whitfield 1983; Russell 1999). Figure 4.x below illustrates the variations in the water quality parameters for the Wilderness and Swartvlei Lake systems.

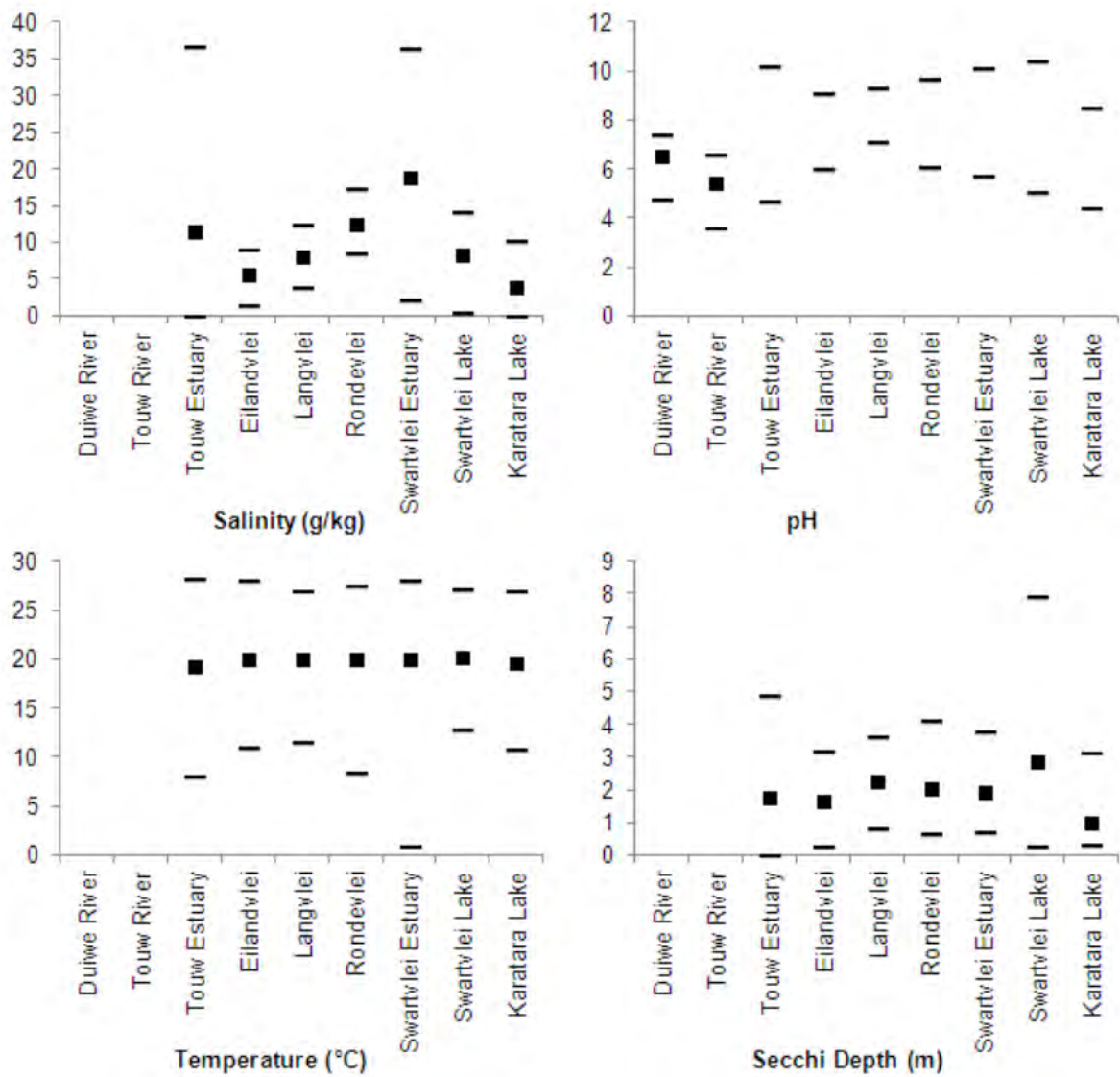


Figure 4.4 Water quality indicators for the Wilderness and Swartvlei Lake systems from 1991 to 1997 (dash = extreme values; square = mean values) (Russel 1999)

Eilandvlei is characterised by an island of dune rock near the middle; Martin (1962) proposes that this island was formerly part of the middle dune cordon, becoming separated by long term erosion – either

as a result of fluctuating sea levels, or by the Kleinkeur River. Furthermore, Martin (1962) found evidence suggesting that the landward side, or northern shore, of the lake represents an earlier coastal shoreline, indicating the occurrence of drastically higher sea levels than observed at present.



Figure 4.5 Eilandvlei, view from the jetty towards the island

In October 2011 the sediment core EV1.11 was recovered at 33°59'23.10" S, 22°38'17.60" E with the final core length being approximately 1.54m. This location was chosen after seismic profiling of the lake (Figure 4.6) indicated thick depositional units at the site.



Figure 4.6 The seismic profile for Eilandvlei indicating the location of the EV1.11 core (Courtesy Kasper 2013; Kirsten 2014)

After retrieval of the core (as described below), mixing was shown for the top 10cm – this section was consequently not included for analysis.

4.3.2. Vibracoring

In order to extract this sediment sequence from Eilandvlei, a vibracorer was used. The vibracorer methodology employed here, was developed by Baxter (1996), modified after Lanesky et al. (1979, as cited by Baxter 1996), and provides access to a range of depositional settings such as lacustrine, fluvial and estuarine environments (Baxter 1996). Baxter (1996) does, however, note a prerequisite for the successful use of the vibracorer; that is that the sediments should be unconsolidated and water-saturated.

For a detailed description of the vibracorer methodology please see Baxter (1996), but in essence sediments are drawn into aluminium tubing, 6m in length, after being fluidised by high frequency vibrations produced by the vibracorer (Fig 4.3). The tube(s) are then retrieved using a winch and tripod. Subsequent to the retrieval of the sediment cores, they are sealed and labelled.

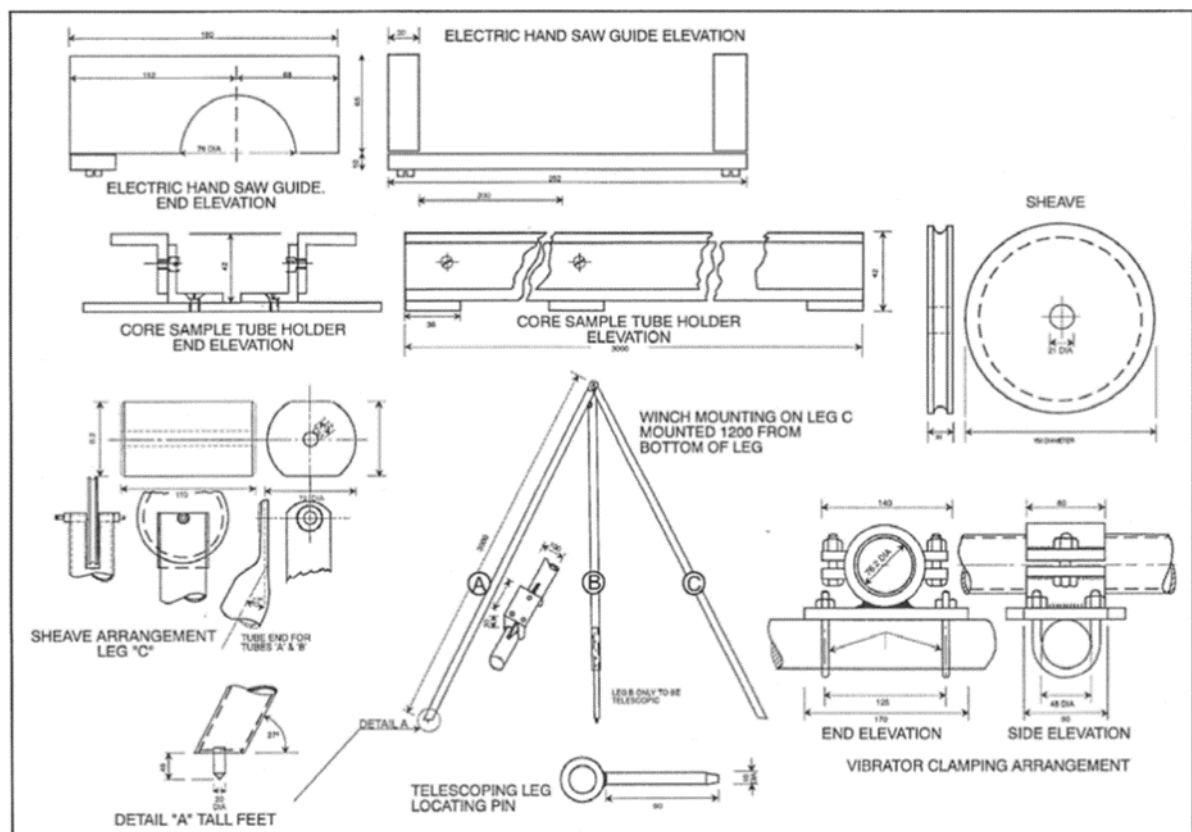


Figure 4.7 The Vibracorer system (Baxter 1996)

4.4. Obtaining Proxy Material

4.4.1. Sediment Sampling

At the laboratory, the core was split laterally and described by colour using Munsell colour notation. As the stratigraphy of the core is relatively uniform, subsampling was done at a resolution of 2cm, where possible, throughout the length of the core. Samples with appropriate organic content was also obtained at this stage to be sent off for radiocarbon dating. The areas identified for subsampling were cleaned and a scalpel was then used to extract these samples – approximately 2g of sediment per sample. Following this, the samples were placed in air-tight plastic vials ready for the extraction and concentration phase.

4.4.2. Laboratory Procedures and Chemical Treatment

The intended aim of the extraction and concentration of pollen from sediment samples is the production of a pollen-rich sample, allowing for the accurate identification and counting of pollen grains present in the sample (Moore et al. 1991). This processing procedure also preserves charcoal particles present within the sample, allowing the simultaneous analysis of both proxies.

To obtain these pollen-rich samples, several chemical treatments are used aimed, at the elimination, or disintegration and dissolution of the non-pollen matrix present in the sample (Moore et al. 1991). The variety of chemical treatments used depends on the extraneous materials present in the sample. For example, lacustrine sediments, in general, contain either siliceous or calcareous mineral material (Moore et al. 1991), each requiring different treatments. Standard palynological methods were employed as per Faegri & Iversen (1989) and Moore et al. (1991) with adaptations from Quick (2013) applied where necessary.

The following chemical treatments were employed: (For the detailed laboratory procedure please see Appendix A)

Firstly, sodium hydroxide (NaOH) is added to the sample for the purpose of disintegrating the pollen matrix, as well as to dissolve humic acids present in the sample. The use of potassium hydroxide (KOH) is also suggest for this step, but Duncan (2006) obtained improved results through the use of NaOH. Preceding any further chemical treatments, a known amount of the exotic pollen *Lycopodium* is added to each sample to aid with the determination of absolute pollen counts and concentrations. Next, hydrochloric acid (HCl) is added in order to eliminate any calcium carbonate (CaCO₃) present in the matrix, followed by Potassium hydroxide (KOH) treatments to remove humic acids. Zinc chloride

(ZnCl₂), with a specific gravity of 1.88, is then used for the heavy liquid mineral separation step isolating pollen grains from the mineral portion still left within the sample. Due to the siliceous nature of the samples, treatment with hydrofluoric acid (HF) is required. Through the use of HF, silica in the sample is removed which might obscure pollen grains when the sample is finally mounted. Acetolysis was further done to remove any cellulose and/or organic matter still present in the sample. Lastly the samples are mounted on slides using the aqueous mounting agent *Aquatex*.

4.4.3. Counting and Identification

The counting and identification of pollen, and charcoal, are very labour intensive. As such, the accepted, or suggested, pollen count per level to be deemed statistically significant is 300 grains. It is however recommended that the 300 benchmark be used as a minimum value instead while rather aiming for a count of 500 grains per level (Bennett & Willis 2001).

Following this recommendation, counts of 500 grains or three slides, whichever came first, per level were performed. A Zeiss Axiostar Plus microscope was used with a magnification of 400x as suggested for the effective and accurate counting and identification of pollen grains (Moore et al. 1991). A magnification of 1000x was further used for the identification of more cumbersome taxa or where pollen grains were possibly damaged complicating identification. To aid in identification, the pollen reference slide- and photograph collection from the University of Cape Town Environmental and Geographical Science department was used.

Absolute counting methods were employed, aided by the addition of the exotic marker *Lycopodium*. To assist in the counting process, the software programme Polycounter version 2.5.3 (Nakagawa 2007) was used. Through the absolute counting method, the actual number of pollen grains present per sample unit can be determined without having to count every grain (Moore et al. 1991). This is possible due to the relationship between the counted and added number of exotic markers, which is then used to approximate the relationship between the counted and actual number of pollen grains present in the sample (Moore et al. 1991). The following formula from Faegri and Iversen (1989) was used:

$$\text{Total Fossil Pollen} = \frac{\text{fossil pollen counted} \times \text{total number of markers}}{\text{exotic markers counted}}$$

As charcoal particles are also preserved during pollen preparation, and thus are also present on the slides, they are counted in conjunction with pollen grains. Through statistical studies, Tinner and Hu (2003) found it unnecessary to painstakingly measure or estimate the area of charcoal particles in pollen slides, as per the point-count method of Clark (1982). Instead, it is suggested that counts of particle numbers are adequate as it is the most dependable and time efficient method (Tinner & Hu 2003; Conedera et al. 2009). As such, the particle count method was employed with only fragments > 75 μm^2 (or $\sim 10 \mu\text{m}$ in length), black in colour, opaque and angular counted.

4.4.4. Data Representation and Analysis

From the results obtained through pollen and charcoal analysis, diagrams were drawn up using the software package Tilia (version 1.7.16) (Grimm 1997). The data were further plotted against the calibrated radiocarbon dates as well as the stratigraphic layers (Faegri & Iversen 1989) to produce standard diagrams. Zonation of the diagrams was achieved with the use of cluster analysis by the application of CONISS (Constrained Incremental Sum of Squares) (Grimm 1987) - a function of the Tilia software program.

To further aid in the interpretation of the results, several ordination techniques are available from which possible patterns within the data can be explored. Here, Detrended correspondence analysis (DCA) was applied to the data first, to determine the length of the environmental gradient. This was done using the program CANOCO for Windows (version 4.51) and CanoDraw for Windows (version 4.1) (Ter Braak & Smilauer 1997). Leps and Smilauer (2003) suggest the use of Principal Component Analysis (PCA) where the gradient length is indicated to be less than three. For gradient lengths exceeding four, unimodal methods (e.g. DCA, Correspondence Analysis) are recommended (Leps & Smilauer 2003).

4.4.5. Radiocarbon Dating and Calibration

Samples deemed suitable, i.e. samples containing sufficient organic matter, were sent to Beta Analytic Inc., Miami, USA for radiocarbon dating. Here accelerator mass spectrometry (AMS) is employed. This method counts the amount of ^{14}C atoms in relation to the stable carbon isotopes ^{13}C and ^{12}C , providing a ratio which is then compared to samples with a standard known ^{14}C content to give the radiocarbon age.

Upon receiving the radiocarbon age results, calibration of these ages are necessary. Calibration converts radiocarbon ages to calendar dates and compensates for the invalid assumption of no variation in atmospheric ^{14}C , made during the calculation of radiocarbon ages (Stuiver & Polach 1977).

The ages were calibrated using the CALIB radiocarbon calibration program (Stuiver et al. 2014) and the calibration data set SHCal13 (Hogg et al. 2013). CLAM (Blaauw 2010) source code was used within the R statistical platform to obtain the age-depth model.

4.5. Conclusion

While outlining the methodological approach for the analysis of palaeo-proxy material, this chapter further highlighted several problems that are still present in the field of palynology – most notably those relating to taphonomy and taxonomy. The results from Eilandvlei, obtained as described here, are provided in the following chapter.

5. Palaeoenvironmental Results from Eilandvlei

5.1. Introduction

In the previous chapter the methodological approach to obtain palaeoenvironmental evidence was discussed. The results attained through these methods are now presented for EV1.11.

5.2. Chronology and Sediment Accumulation

Six radiocarbon ages were obtained for EV1.11. These results, given in Table 5.1, show the core to incorporate the time period from c. 3 880 cal yr BP to present, setting the late Holocene as the period of investigation.

Table 5.1. Radiocarbon and calibrated ages for the EV1.11 core

Sample	Depth (cm)	14C age yr BP	Cal BP	Calibration	95.4 % (2 σ) cal BP range			Median probability
					Lower	Upper	Relative probability	
EV1.11 - 2	20		-7					
EV1.11 - 3	40	470		SHCal13	340	352	0.031	497
					451	526	0.969	
EV1.11 - 7	78	1290		SHCal13	1072	1192	0.649	1171
					1206	1268	0.351	
EV1.11 - 4	102	2140		SHCal13	2002	2154	0.989	2074
					2276	2287	0.011	
EV1.11 - 6	134	3300		SHCal13	3397	3568	1	3485
EV1.11 - 5	151	3620		SHCal13	3726	3751	0.039	3880
					3791	3794	0.055	
					3820	3980	0.907	

The age depth model, Figure 5.1., indicates continuous deposition of sediment. For the first 135cm of EV1.11, 154 to 20cm, an average sedimentation rate of 0.38mm/yr is observed. A significant increase in this accumulation rate, to 3.77 mm/yr, is then seen at the top of the core from 19cm.

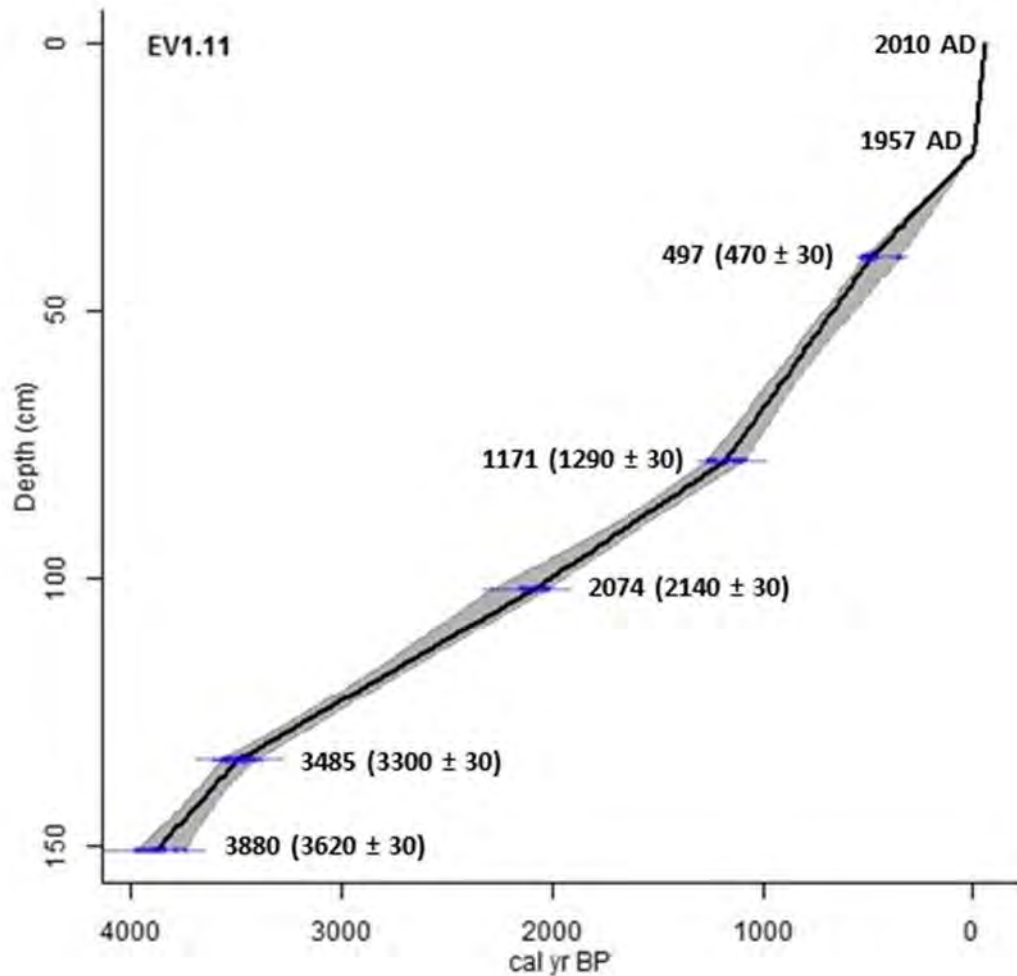


Figure 5.1. The age-depth model for EV1.11, produced in *CLAM* (Blaauw 2010)

5.3. Stratigraphy

The core extracted from Eilandvlei, EV1.11 shown in Figure 5.2, has a total length of 154 cm. The stratigraphy appears to be fairly homogeneous, although the sediments towards the base of the core are distinctly darker in colour.

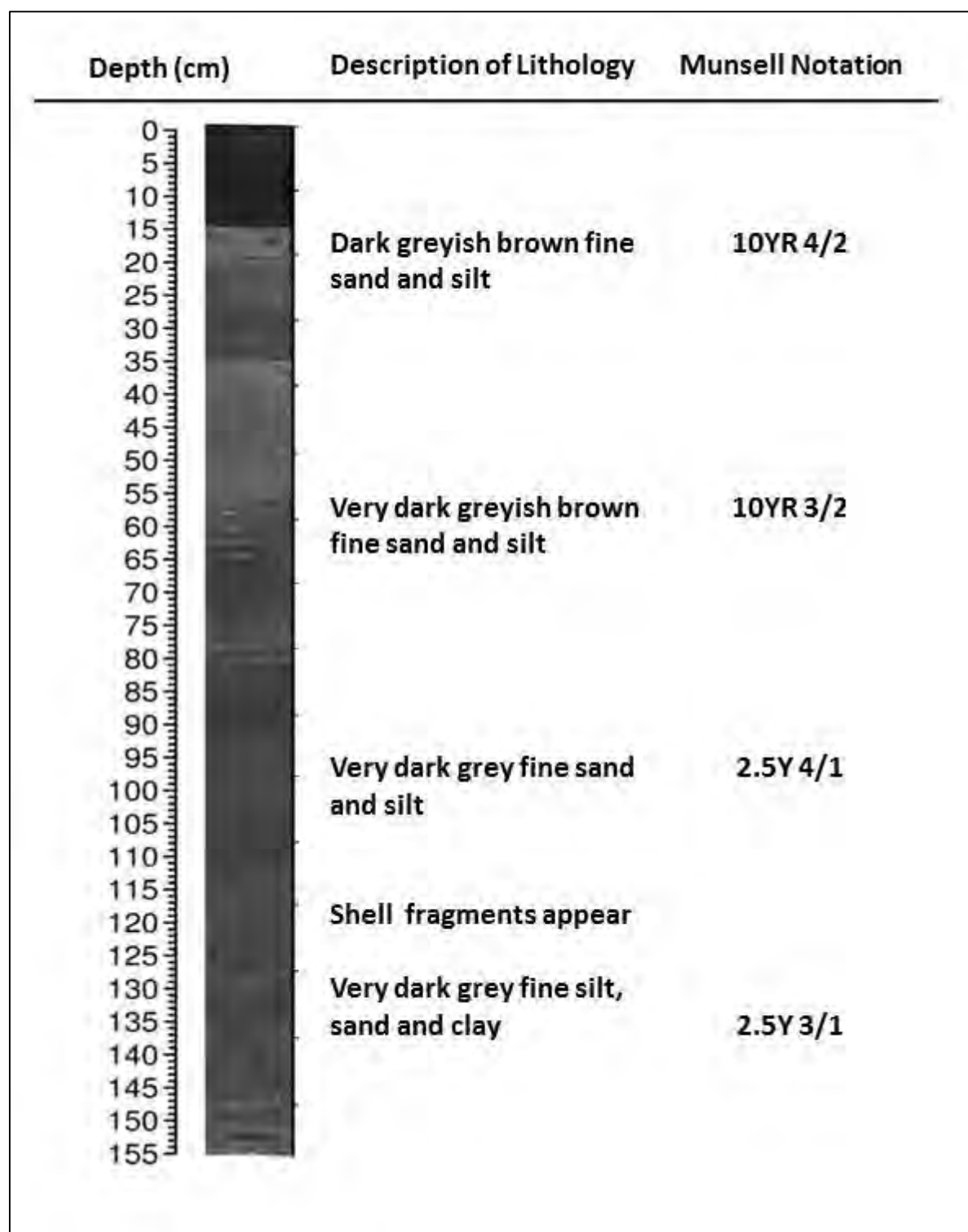


Figure 5.2. Stratigraphic description of EV1.11 according to the Munsell Colour notation

5.4. Pollen and Microscopic Charcoal Analyses

In general an increase in the pollen concentration is observed towards the top of the core. Values range from 6.26×10^3 grains/g near the base, to a maximum of 7.86×10^4 grains/g around the top of the sequence, with an average concentration of 2.72×10^4 grains/g.

5.4.1. Pollen Assemblage Zones

Through the use of cluster analysis, five distinct pollen assemblage zones were identified, starting from the base, zone EV1.11A to EV1.11E as indicated in Figure 5.3 below.

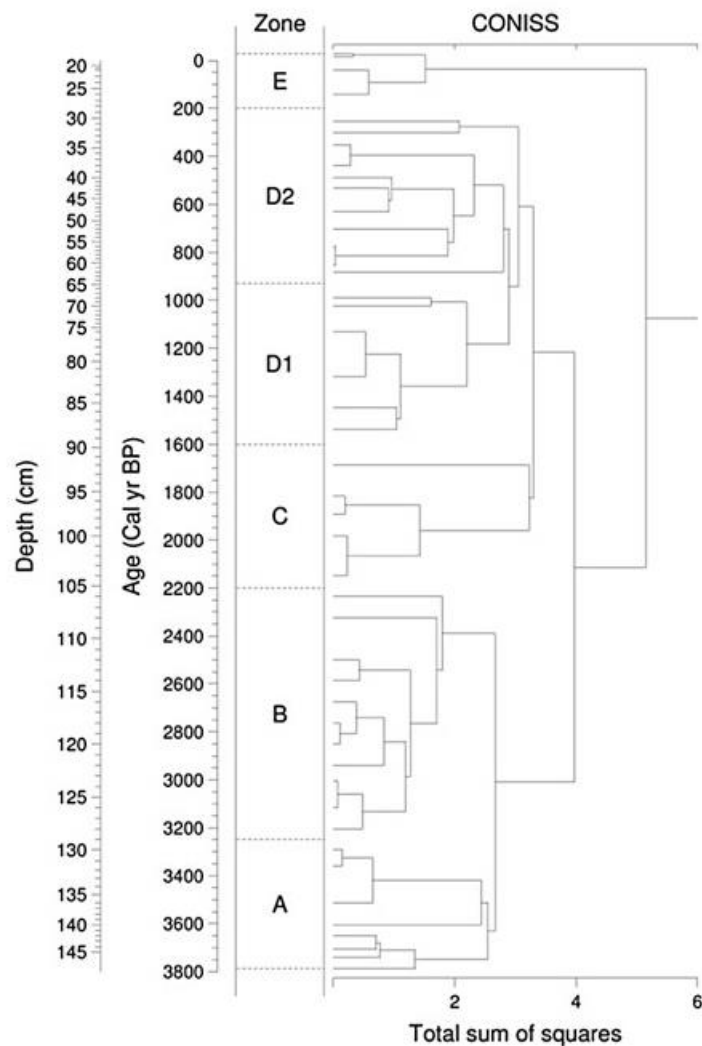


Figure 5.3. The pollen assemblage zones for EV1.11

The results obtained from the pollen and charcoal analyses are discussed below, according to the assemblage zones. This is accompanied by the relative percentage diagrams, Figures 5.4-1 to 5.4-3, Figures 5.5-1 and 5.5-2, and Figure 5.6 and 5.7.

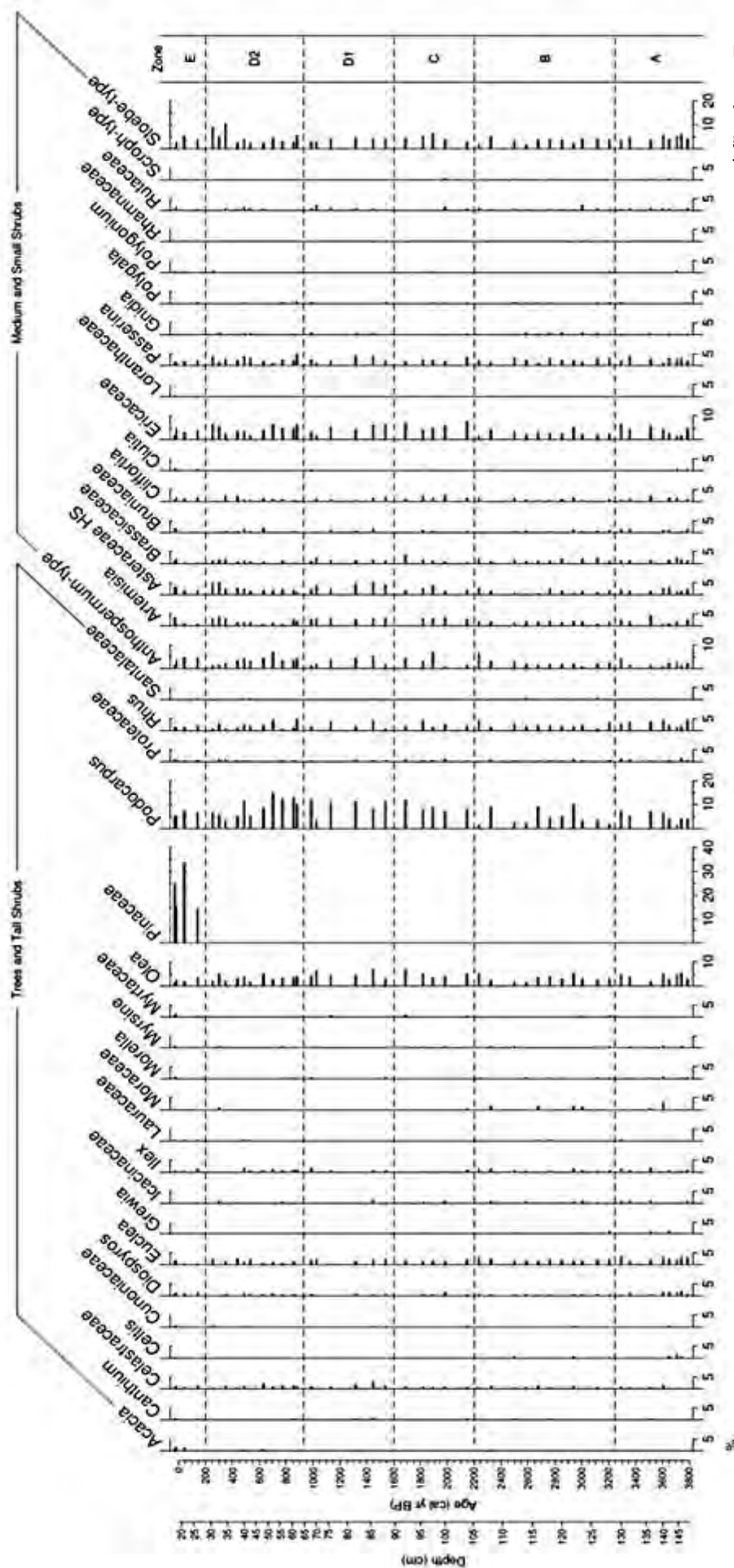


Figure 5.4-1. The relative pollen percentage diagram for EV1.11, with taxa grouped according to general growth forms: 1 of 3: Trees and Tall shrubs, and Medium and Small shrubs

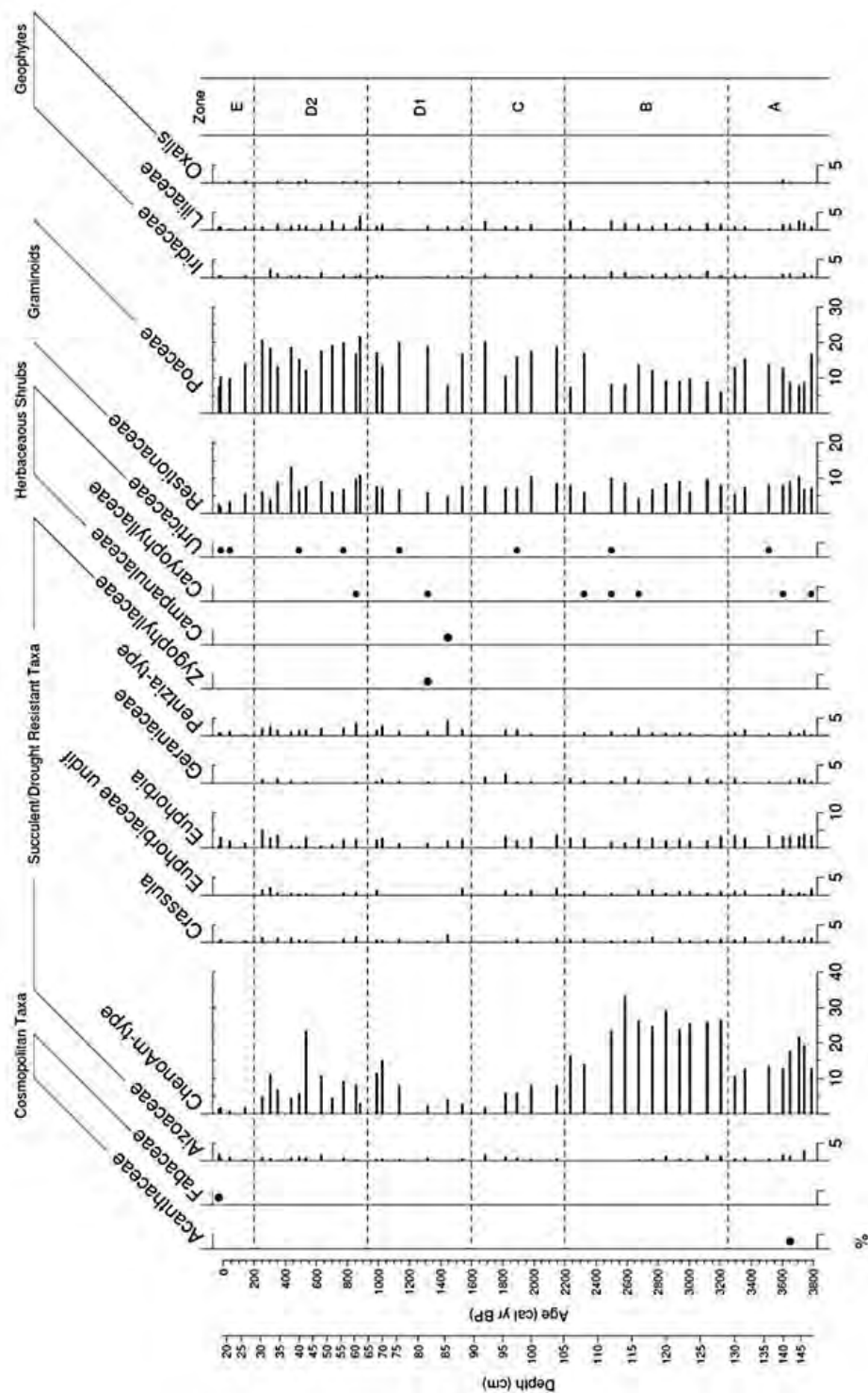


Figure 5.4-2. The relative pollen percentage diagram for EV1.11, with taxa grouped according to general growth forms: 2 of 3: Cosmopolitan taxa to Geophytes

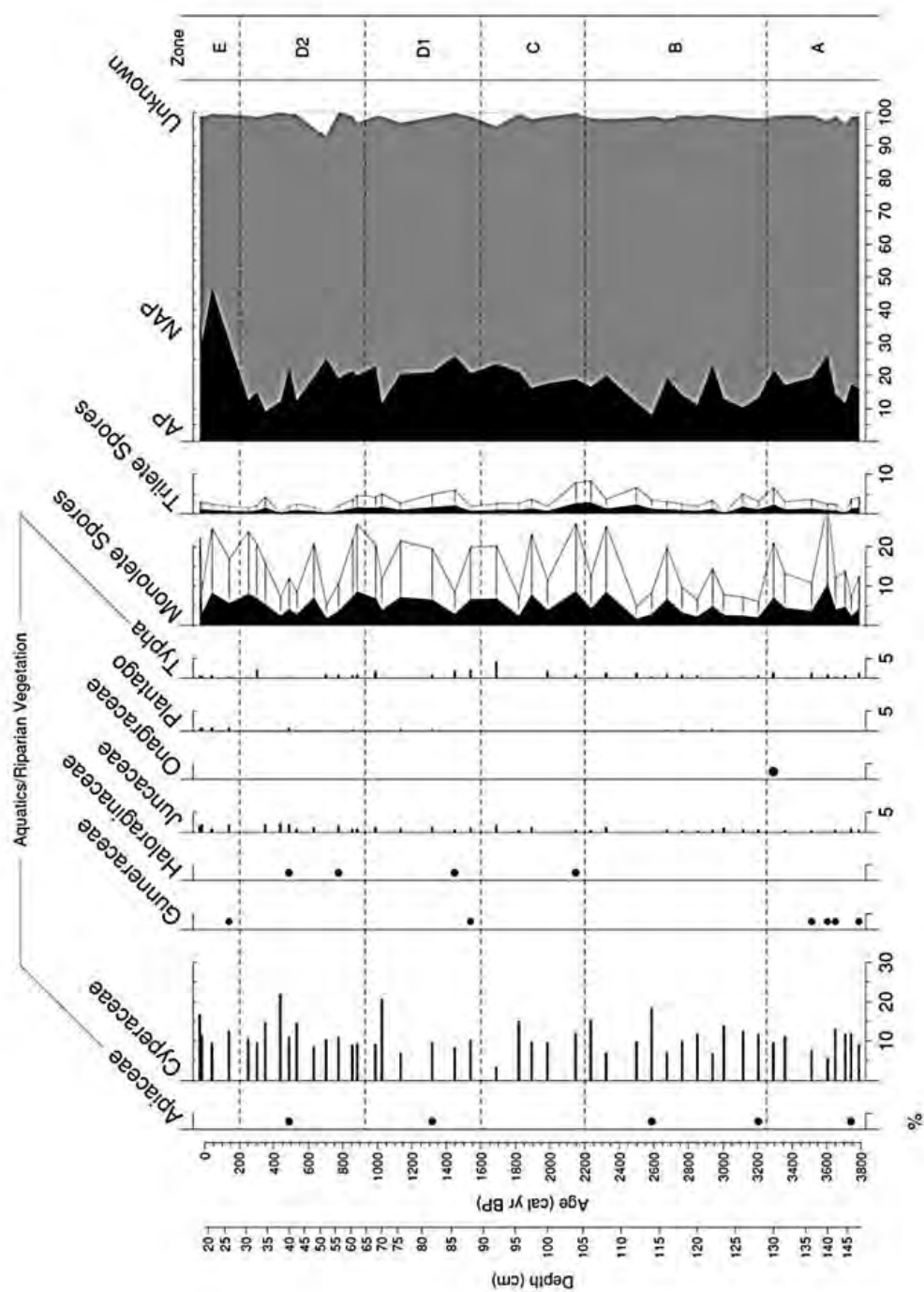


Figure 5.4-3. The relative pollen percentage diagram for EV1.11, with taxa grouped according to general growth forms: 3 of 3: Aquatic/Riparian Vegetation, mono- and trilete spores and the Arboreal (AP) and Non arboreal (NAP) percentages

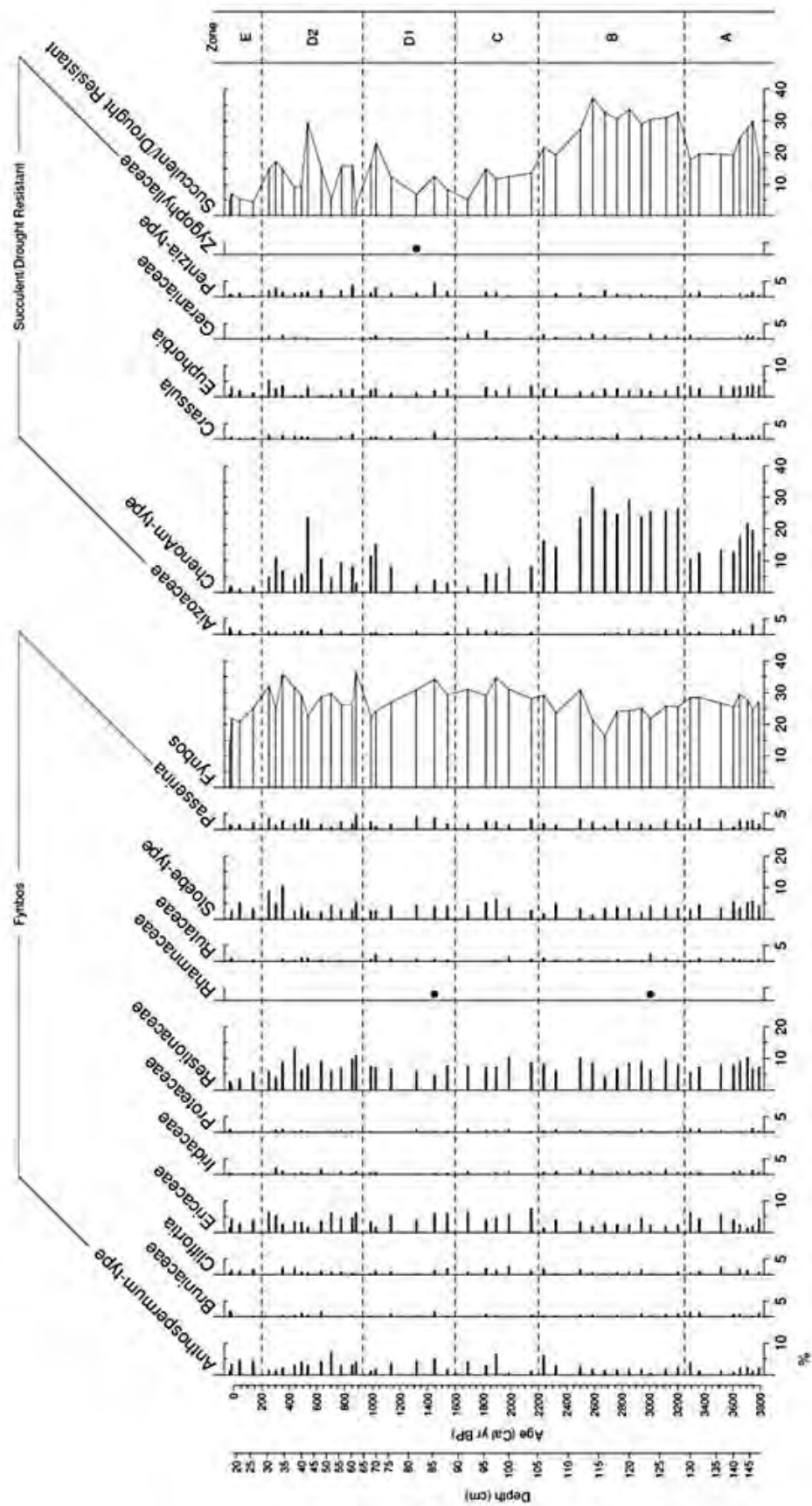


Figure 5.5-1. The EV1.11 relative percentage pollen diagram, with selected taxa grouped according to ecological affinity: 1 of 2: Fynbos and Succulent/Drought resistant groups

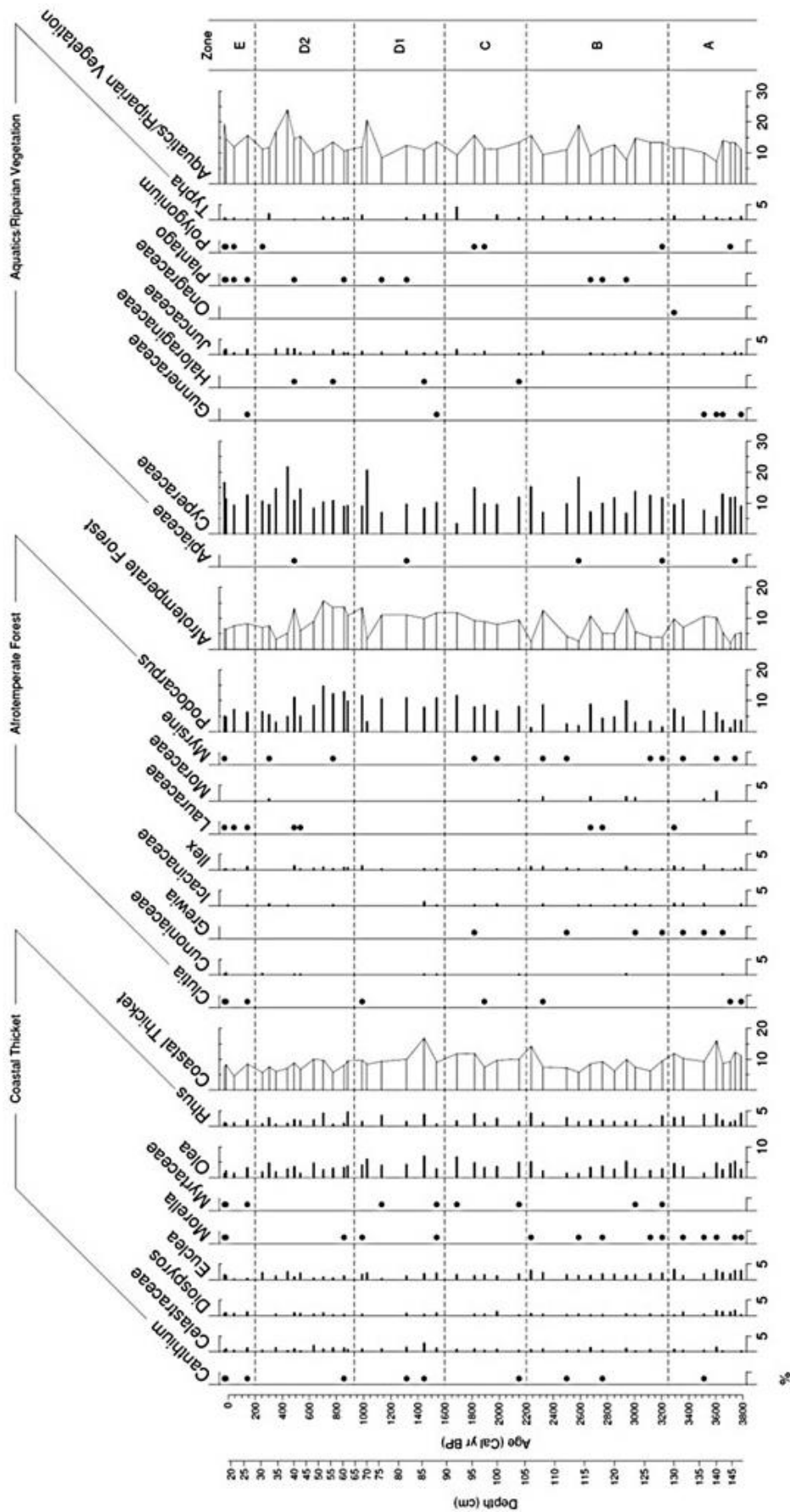


Figure 5.5-2. The EV1.11 relative percentage pollen diagram, with selected taxa grouped according to ecological affinity: 2 of 2: Coastal Thicket, Afrotropical Forest and Aquatic/Riparian vegetation groups

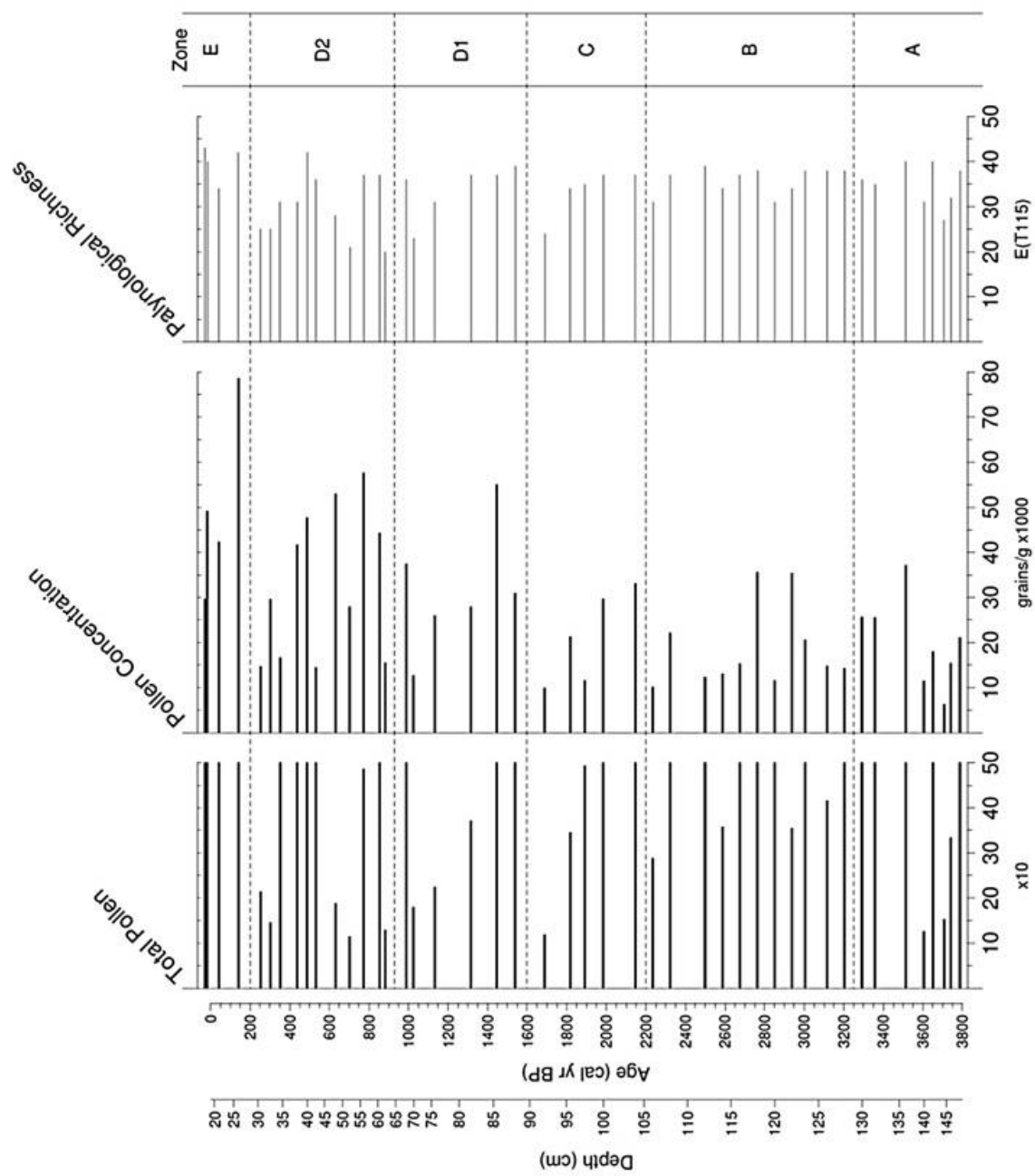


Figure 5.6. The total pollen count, pollen concentration and palynological richness for EV1.11

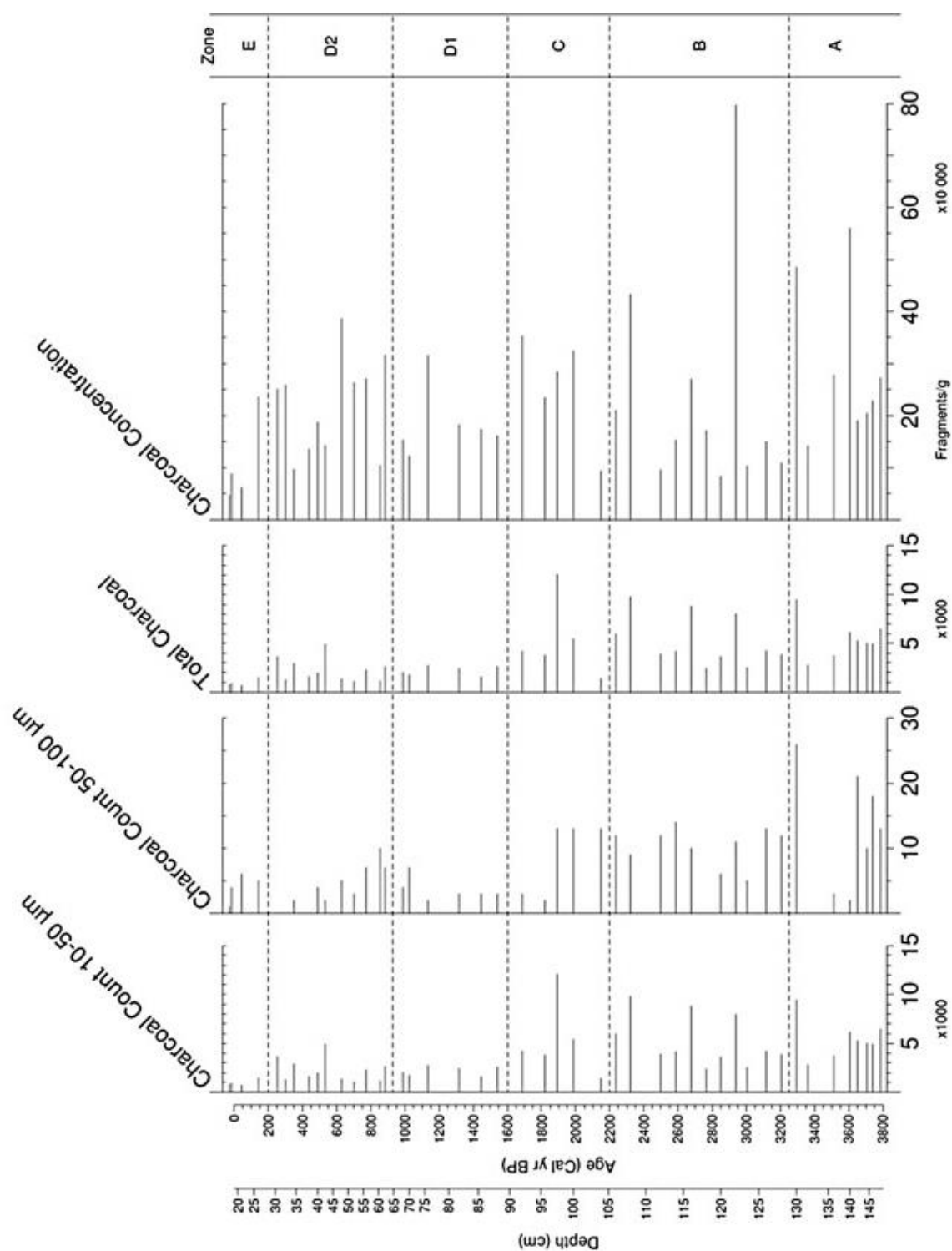


Figure 5.7. The EV1.11 microscopic charcoal analysis results: charcoal counts for the two size classes, the total charcoal sum and charcoal concentration

Zone EV1.11-A: 148 – 130 cm; c. 3790 – c. 3290 cal yr BP

This zone, representing the basal part of the core, is dominated by the presence of Restionaceae, *ChenoAm*-type, *Podocarpus*, Poaceae and Cyperaceae pollen.

Poaceae shows a general increasing trend throughout the zone from 146cm (c. 3740 cal yr BP), with a significant peak (16.6%) observed at the base of the core, 148cm (c. 3790 cal yr BP).

Ericaceae (1.3%) reaches its minimum value for the entire sequence towards the base of the core at 144.5cm (c. 3700 cal yr BP). This low proportion of Ericaceae pollen coincides with the highest values of Restionaceae (10.5%), *ChenoAm*-type (21.6%) and Liliaceae (2.6%) achieved in the zone. An increase in the percentages of the Fynbos elements *Anthospermum*-type (2.6%), *Stoebe*-type (5.2%) and *Passerina* (2.6%) are also evident at this time. This low presence of Ericaceae also coincides with low proportions of *Cliffortia* (1.3%) and Asteraceae HS (1.3%) as well as *Rhus* (1.3%) as it also reaches its lowest value for the zone here. Afrotropical Forest elements are virtually absent at this point except for *Podocarpus* (1.3%) at its lowest percentage for the sequence, and indeed, Afrotropical Forest elements as a whole are at their lowest levels for the assemblage.

Around the middle of the zone, at 140cm (c. 3600 cal yr BP), *Passerina* (0.8%) displays its minimum value within the sequence. Concurrently, *Morella* (0.8%), two Afrotropical Forest elements, Moraceae (3.2%) and Myrsine (0.8%), and the aquatic *Gunneraceae* (0.8%) reach maximum proportions for the whole of the sequence.

At the top of the zone (130cm, c. 3290 cal yr BP) Onagraceae is present for the only time within the sequence. This coincides with the first appearance of Lauraceae (0.4%) in the zone, while Ericaceae (6.4%) reaches its maximum value for the zone here. This is also where *Anthospermum*-type (4%), *Podocarpus* (1.2%) and *Typha* (1.4%) achieve their highest proportions within the zone, while Proteaceae (1.2%) and *Euclea* (3.4%) reach their maximum proportions for the whole sequence. At the same time, Restionaceae (5.6%), *Stoebe*-type (3.2%) and *ChenoAm*-type (10.6%) show minimum proportions within the zone.

Two major peaks in the charcoal concentration are evident at 140cm (c. 3600 cal yr BP) and 130cm (c. 3290 cal yr BP). The larger charcoal particle group, 50-100 µm, reaches a maximum at the top of the zone at 130cm (c. 3200 cal yr BP).

The monolete spores group also show two significant peaks around the same time, displaying maximum levels for the sequence at the first peak (140cm, c. 3600 cal yr BP). A peak in trilete spore levels is observed at the bottom of the zone (148cm, c. 3790 cal yr BP) while an increasing trend is

shown towards the top of the zone after an absence from the assemblage at 144.5cm (c. 3700 cal yr BP).

Zone EV1.11-B: 128 – 106cm; c. 3200 – c. 2230 cal yr BP

This is the largest of the pollen assemblage zones encompassing a time span of around 1000 years. As in Zone EV1.11-A, the most prominent taxa are Restionaceae, *Podocarpus*, Cyperaceae, *ChenoAm*-type and Poaceae.

Restionaceae appears to follow a cyclical trend throughout the zone. The same is evident for both *Stoebe*-type and *Passerina* although the observed peaks and troughs are not contemporaneous. Poaceae exhibits a general increasing trend starting with its minimum value for the entire sequence (6%) at the bottom of the zone at 128cm (c. 3200 cal yr BP) and two noteworthy peaks observed at 116cm (13.8%) and 108cm (17%) respectively. Rhamnaceae makes its first appearance in the sequence around the bottom of the zone at 123.5cm (c. 3000 cal yr BP). Aizoaceae is completely absent from the upper part of the zone, only appearing at 116cm (c. 2670 cal yr BP). *Clutia* is seen, for the first time in Zone B, near the top of the zone at 108 cm (c. 2320 cal yr BP).

Near the base of this zone, at 122cm (c. 2940 cal yr BP), Bruniaceae (1.1%), Ericaceae (4.8%), Proteaceae (0.8%), *Olea* (5.4%), *Ilex* (1.1%) and *Podocarpus* (10.1%) display their highest percentages within the zone, while Cyperaceae values are at a minimum (6.8%). Cunoniaceae (0.3%) is also present here for the only time within this zone.

It is clearly evident that *ChenoAm*-type pollen is the most prevalent within this zone achieving its highest percentage for the whole sequence (33.2%) at 114cm (c. 2590 cal yr BP). Both Asteraceae HS (0.3%) and *Olea* (1.4%) further display their minimum values within the sequence here, while Proteaceae is completely absent. A significant peak in Cyperaceae (18.4%) is also observed here while both *Stoebe*-type (1.4%) and *Euphorbia* (1.4%) exhibit minimum values for the zone. *Podocarpus* is also present albeit in very low proportions (2%).

The 10-50 µm charcoal count also exhibits a cyclical trend, a pattern that is mirrored by the total charcoal count. Three large peaks are evident for the charcoal concentration with a significantly large peak at 122cm (c. 2940 cal yr BP) – this is indeed the maximum charcoal concentration for the entire sequence at 7.974×10^5 fragments/g. Even though fluctuations in the levels of the monolete spore group are present, a general increasing trend towards the top of the zone is observed. Trilete spore proportions show minor fluctuations with the highest value for the sequence at 106cm (c. 2230 cal yr BP), the top of the zone.

Zone EV1.11-C: 104 – 92cm; c. 2150 – c. 1690 cal yr BP

This is one of the shorter zones of the assemblage, characterized generally by minor fluctuations in the levels of the most dominant taxa, with only Poaceae, Cyperaceae and *Anthospermum*-type displaying major variations.

Podocarpus exhibits an increasing trend throughout the zone while *ChenoAm*-type pollen percentages decrease towards the top of Zone C. Minor fluctuations in Ericaceae and *Olea* percentages are evident whereas *Stoebe*-type increases towards the top of the zone and *Euphorbia* shows a slight reduction. *Morella* and Lauraceae, as well as the aquatic elements, *Gunneraceae*, *Plantago* and Apiaceae, are completely absent from this zone.

At 104cm (c. 2150 cal yr BP), around the bottom of the zone, Ericaceae (7.6%) pollen reaches its maximum value for the whole sequence. Concurrently, two other Fynbos elements, *Stoebe*-type and *Passerina* are at their lowest (2.8%) and highest (3.4%) levels within the zone respectively. *Canthium* (0.4%), Cunoniaceae (0.2%) and Moreaceae (0.4%) are only present at the base of Zone C. Notable, Haloraginataceae (0.2%) is observed for the first time within the sequence here.

Stoebe-type levels (6.3%), as well as elements of the Fynbos group as a whole, achieve maximum percentage values at 97.5cm (c. 1890 cal yr BP). A distinctive peak in *Anthospermum*-type pollen (6.5%) is also observed here, while the taxa classified as coastal thicket, *Olea* (3.4%) and *Rhus* (1.2%), are at minimum proportions for the zone. Afrotemperate elements are virtually absent here except for the presence of *Clutia* (0.2%) – for the only time within the zone – and *Podocarpus* (8.7%).

At the top of the zone (92cm, c. 1690 cal yr BP), Cyperaceae values (3.4%) are at a minimum for the sequence while *Typha* (4.2%) is at its highest value. Significantly, *Euphorbia* is missing at this time – an absence that is only observed twice within the entire sequence. Myrtaceae (0.8%) and *Olea* (6.7%), representing coastal thicket, display maximum proportions for the zone here. This is also the case for Aizoaceae (1.7%), although levels are relatively constant throughout the zone. *Podocarpus* reaches its highest level (11.8%) within the zone at this time, while all other Afrotemperate components are absent. Poaceae levels achieve a notable peak (20.2%) at the top of the zone.

Both the total charcoal count and the 10-50 µm size class achieve their highest values for the sequence in the middle of the zone, at 97.5cm (c. 1890 cal yr BP). The 50-100 µm size class shows a significant decrease at this point. The monolete spore group displays a cyclical trend throughout the zone while the trilete spores exhibit minor fluctuations with a relatively large peak at the base of the zone.

Zone EV1.11-D: 88 – 30.5cm; c. 1540 – c. 250 cal yr BP

Zone EV1.11-D, being the largest of the assemblage zones spanning the time period c. 1540 to c. 250 cal yr BP, is divided further into two subzone, D1 and D2.

Zone EV1.11-D1: 88 – 68cm; c. 1540 – 990 cal yr BP

Zone EV1.11-D1, incorporating the basal part of Zone D, is characterised by a general increase of Fynbos throughout the zone, while Succulent/Drought Resistant taxa display a decreasing trend.

At 85.5cm (c. 1450 cal yr BP), Rhamnaceae (0.2%) is present for the only the second time within the entire sequence and Haloragaceae (0.2%) make a first appearance within the zone. The Fynbos elements, *Anthospermum*-type (5%), Bruniaceae (1.6%) and Ericaceae (6%) are present at maximum levels within the zone, whereas Restionaceae (4.8%) reaches its lowest value within the zone here. *Crassula* (2.2%) and Icacinaceae (1.4%) proportions are at their highest for the sequence, with the coastal thicket components *Canthium* (0.8%), Celastraceae (2.8%) and *Olea* (7%) also expressing maximum values within the sequence here.

Zygophyllaceae makes its only appearance in the sequence at 82cm (c. 1320 cal yr BP). Simultaneously, *ChenoAm*-type (2.2%) percentages are at their lowers within the zone, while other Succulent/Drought Resistant taxa are present only in small proportions. The only Afrotemperate element present at this time is *Podocarpus* (11.1%), at its highest level within the zone, with Apiaceae (0.3%) present for the only time within the zone.

In the middle of the zone, at 76cm (c. 1130 cal yr BP), Poaceae (20.1%) displays its maximum proportions within the zone. Higher levels of *Podocarpus* (10.7%) are also seen, with *Ilex* (0.4%) the only other Afrotemperate Forest element present.

Towards the top of the zone (70cm, c. 1030 cal yr BP) significant peaks in both *ChenoAm*-type (15%) and *Olea* (6.1%) are observed. Notably, for the only within the sequence, *Rhus* is absent. A major peak in Cyperaceae (20.6%) is seen here; while a concurrent, noteworthy dip in *Podocarpus* (3.3%) proportions is observed.

Charcoal counts remain relatively stable throughout the zone with only the 50-100 µm class showing a slight peak towards the top of the zone (70cm, c. 1030 cal yr BP), while a significant peak in charcoal concentrations is observed at 76cm (c. 1130 cal yr BP). The monolete spore group again displays a cyclical trend with the trilete spore proportions remaining relatively constant.

Zone EV1.11-D2: 62 – 30.5cm; c. 880 – c. 250 cal yr BP

Zone EV1.11-D2, the upper part of Zone D, is the second largest zone spanning around 600 years. Significant peaks and troughs are evident amongst the most prominent taxa within the zone - Restionaceae, *ChenoAm*-type, *Podocarpus*, Poaceae and Cyperaceae – with the minor components of the assemblage exhibiting cyclical trends.

At 62cm (c. 880 cal yr BP), near the bottom of the zone, increased proportions of the Fynbos elements Ericaceae (6.2%), Liliaceae (3.9%), Restionaceae (10.9%), *Stoebe*-type (5.4%) and *Passerina* (4.7%) are observed. This coincides with very low values for *ChenoAm*-type (3.1%) while all other Succulent/Drought resistant elements are absent. Both *Rhus* (4.7%) and Poaceae (21.7%) achieve their highest values for the sequence here.

Both *Anthospermum*-type (7%) and *Podocarpus* (14.8%) reach maximum proportions for the sequence at 52cm (c. 700 cal yr BP). *Ilex* (0.9%) is the only other Afrotemperate Forest element present at this time. Concurrently, slight peaks are observed in the Fynbos elements Ericaceae (6.1%) and *Stoebe*-type (4.3%) while Restionaceae (6.1%) displays lower values. Increased levels in Poaceae (19.1%) and *Rhus* (4.3%) are also observed.

In the middle of the zone (42.5cm, c. 530 cal yr BP), a major peak in *ChenoAm*-type pollen is observed. This is indeed also the highest level of *ChenoAm*-type pollen within this zone. Coincident with this peak, Poaceae (12.2%) displays its minimum value within the zone. A slight peak in Cyperaceae (14.6%) is visible after relatively constant levels for the bottom part of the zone.

Both the Aquatic/Riparian components Cyperaceae (21.8%) and Juncaceae (2.2%) are at their highest levels for the sequence at 38cm (c. 440 cal yr BP). Restionaceae (13.2%) pollen also achieves its maximum value for the sequence at this time together with a large peak in Poaceae (18.6%).

Stoebe-type proportions (10.4%) reach a maximum at 34.5cm (c. 350 cal yr BP) while *Podocarpus*, the only Afrotemperate Forest element present at this time, is at a relatively low percentage (3.2%).

At the top of the zone (30.5cm, c. 250 cal yr BP), *Euphorbia* (5.1%) reaches its maximum value for the sequence. This is accompanied by increased levels of the Fynbos components Ericaceae (6.5%), *Stoebe*-type (8.9%), *Passerina* (3.7%) and Restionaceae (6.1%). Poaceae (20.6%) also exhibits a major peak at this time, while *Polygonium* (0.5%) is present for the only time within the zone.

Large peaks in both the 10-50 µm charcoal particle count and total charcoal count are observed at 42.5cm (c. 530 cal yr BP). The 50-100 µm size class exhibits a decreasing trend towards the top of the zone while charcoal concentrations peak at 48cm (c. 630 cal yr BP). Fluctuations throughout the zone

are observed for both monolete and trilete spores groups, with both displaying increased levels at the bottom and top of the zone.

Zone EV1.11-E: 26 – 12cm; c. 1810 – 1980 AD (c. 140 – c. -30 cal yr BP)

This is the shortest of the zones only covering a time span of 110 years or so. Zone EV1.11-E is further characterised by the presence of the neophyte *Pinus*.

Overall, the Succulent/Drought Resistant group is significantly reduced compared to the rest of the previous zones, with *ChenoAm*-type displaying its minimum value for the sequence (0.8%) around the middle of the zone at 16.5cm (c. 1970 AD).

The Aquatic elements Cyperaceae, Juncaceae, *Plantago*, *Polygonium* and *Typha* are consistently present throughout the zone while *Gunneraceae* is only seen at the start of the zone.

Podocarpus proportions remain relatively constant throughout the zone with only a minor increase, up to 7.2%, observed at around 1910 AD at 22cm. In contrast, *Pinus* percentages are notably higher with large fluctuations between levels with values ranging from 14.2% to 33.6%.

Increased percentages of Poaceae are observed at the top of the zone (26cm, c. 1810 AD). This is accompanied by higher proportions of the coastal thicket elements *Olea* (3.2%) and *Rhus* (2%) as well as increased percentages of Cyperaceae and Restionaceae.

At the top of the zone, representing the time period c. 1980 AD (12cm), *Stoebe*-type (0.12%) and *Olea* (1.4%) proportions are at their lowest percentages for the whole of the sequence, while a relatively large peak in Cyperaceae (16.8%) is observed. Despite the presence of the majority of the fynbos indicator taxa at the top of the zone, the overall proportion of pollen representing the fynbos group is reduced in comparison to the rest of the zone.

The 10-50 µm charcoal size class, total charcoal, as well as charcoal concentrations all decrease towards the top of the zone reaching their lowest levels for the sequence. The larger charcoal fragment count, 50-100 µm, deviates from this pattern displaying constantly increased values throughout the zone except for at the very top. Monolete spores show fluctuating values throughout the zone while proportions for the trilete spore group remain relatively constant with a slight increase towards the top of the zone being observed.

5.5. Statistical Analysis

18 taxa, identified as important ecological indicators (Table 5.2), were selected for further statistical analysis. Detrended correspondence analysis was initially performed first in order to extract underlying gradients within the data, gradient lengths below 3 were indicated for this data set. For gradient lengths between 1.5 and 3 either linear or unimodal ordination methods can be applied (Ter Braak & Prentice, 1988) consequently, principal component analysis (PCA) was performed.

Table 5.2. Pollen taxa selected for statistical analysis

Pollen Taxa	Ecological Affinity
<i>Cliffortia</i> Ericaceae Proteaceae <i>Stoebe-type</i> <i>Passerina</i>	Fynbos
Aizoaceae <i>ChenoAm-type</i> <i>Euphorbia</i> <i>Pentzia-type</i>	Succulent/Drought Resistant
Celastraceae <i>Olea</i> <i>Morella</i> <i>Myrtaceae</i>	Coastal Thicket
<i>Podocarpus</i> Lauraceae Cunoniaceae	Afrotemperate Forest
Gunneraceae Haloragaceae	Aquatic/Riparian

Table 5.3. Principal Component Analysis results showing the first four components and the variance explained by each

Principal Component Axis	1	2	3	4	Total Variance
Eigenvalues	0.172	0.153	0.127	0.112	1
Cumulative percentage variance of species data	17.2	32.5	45.2	56.4	

Table 5.3 indicates that 72.8% of the variance is explained by the first four principal components with the eigenvalue of the first component (PC1), 0.489, being significantly larger than the other three,

constituting almost half of the total variance. PC1 is thus the most important component extracted from the data, while PC2 is added here to present a more comprehensive picture.

Table 5.4. The loadings for the first two principal components, with significant positive loadings in red and significant negative loadings in green

Pollen Taxa	PC1	PC2
<i>ChenoAm</i> -type	0,982	-0,174
<i>Euphorbia</i>	0,415	0,031
<i>Morella</i>	0,142	0,235
<i>Passerina</i>	-0,003	-0,026
Lauraceae	-0,065	0,030
Gunneraceae	-0,069	0,152
Aizoaceae	-0,079	0,061
<i>Pentzia</i> -type	-0,096	-0,375
<i>Olea</i>	-0,114	-0,312
<i>Cliffortia</i>	-0,121	0,305
Proteaceae	-0,128	0,025
<i>Stoebe</i> -type	-0,138	-0,160
Haloraginaceae	-0,217	-0,240
Cunoniaceae	-0,226	0,089
Myrtaceae	-0,306	0,425
Celastraceae	-0,533	-0,349
Ericaceae	-0,611	-0,430
<i>Podocarpus</i>	-0,679	-0,672

ChenoAm-type shows a very large positive loading for PC1. *Euphorbia*, also a component of the succulent/drought resistant group, similarly is associated with a significant positive loading although not on the same scale as that of *ChenoAm*-type.

Significant negative loadings are observed for *Podocarpus*, Ericaceae and Celastraceae, suggesting an interesting dynamic between forest and drought resistant taxa.

Although PC2 only contributes >10% of the total variance, it is still interesting to note that both *Podocarpus* and Ericaceae show significant negative loadings and Myrtaceae, a coastal thicket element, is the only taxon displaying a significant positive loading.

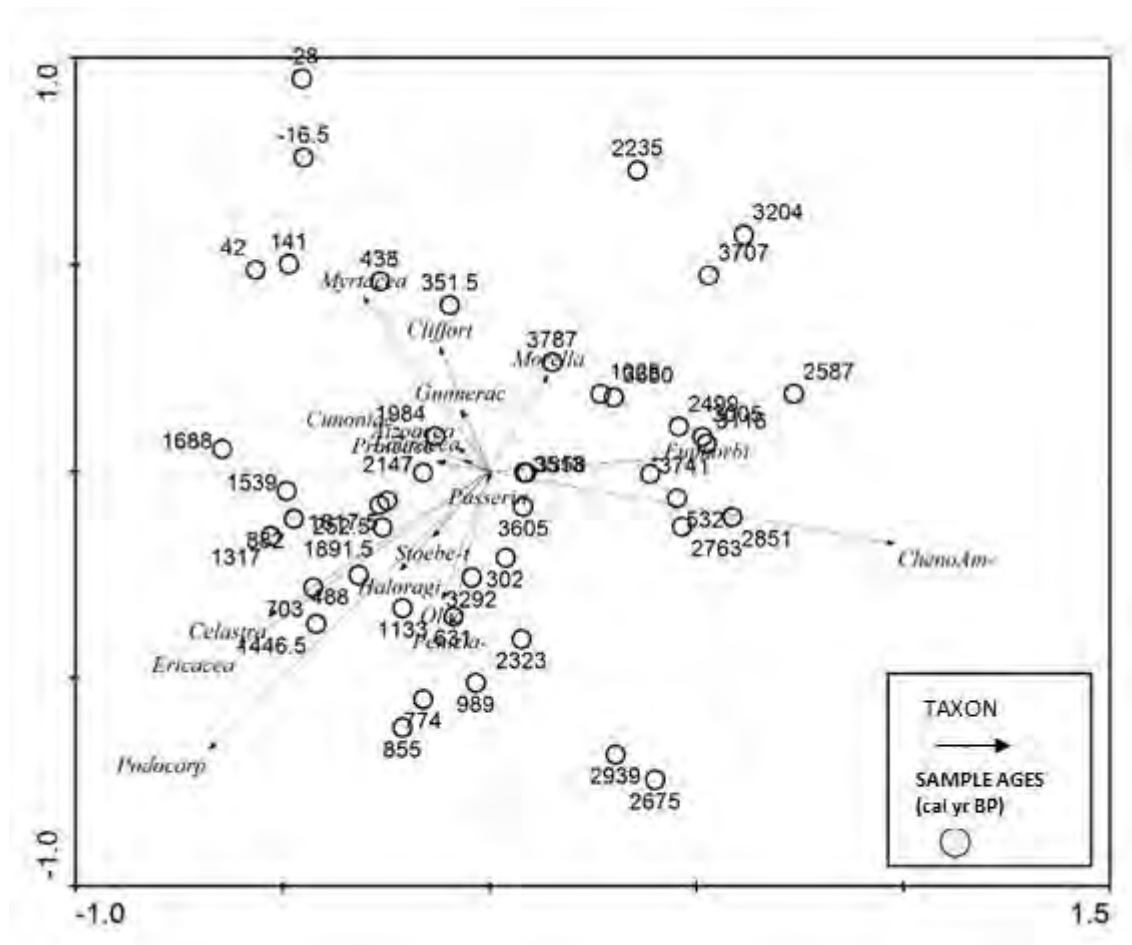


Figure 5.8. Ordination biplot of the first two principal components, with PC1 on the x-axis and PC2 on the y-axis

Figure 5.8 further illustrates the relationship between *ChenoAm*-type and *Podocarpus* showing that they are present on opposite sides of the plot indicating their contrasting component loadings. *ChenoAm*-type also finds itself alone in the bottom right hand quadrant, while *Euphorbia* and *Morella* are the only other taxa present on that side of the plot.

Based on the results from this PCA, it was decided to rerun the analysis excluding *ChenoAm*-type and *Podocarpus*, as the relationship between these two dominant taxa may be obscuring the analysis. These results follow below.

Table 5.5. Results from the second PCA showing the first four components and the variance explained by each

Principal Component Axis	1	2	3	4	Total Variance
Eigenvalues	0.172	0.153	0.127	0.112	1
Cumulative percentage variance of species data	17.2	32.5	45.2	56.4	

The total variance explained by the first four principal components is 56.4%. In contrast to the first PCA, the first component (PC1) only contributes 17.2% to the total variance with the cumulative variance of the first three components barely explaining half of the total variance within the data.

Table 5.6. The principal component loadings for the second analysis, with significant positive loadings in red and significant negative loadings in green

Pollen Taxa	PC1	PC2
Myrtaceae	0,436	-0,070
<i>Morella</i>	0,161	-0,398
Celastraceae	0,095	0,692
Ericaceae	0,024	0,802
<i>Cliffortia</i>	0,024	0,142
Lauraceae	-0,004	-0,139
Gunneraceae	-0,030	0,002
<i>Olea</i>	-0,030	0,330
Haloraginnaceae	-0,074	0,348
Cunoniaceae	-0,110	0,036
Aizoaceae	-0,120	-0,361
<i>Passerina</i>	-0,147	0,189
Proteaceae	-0,286	0,048
<i>Stoebe</i> -type	-0,447	0,323
<i>Euphorbia</i>	-0,656	-0,353
<i>Pentzia</i> -type	-0,820	0,140

For the first principal component, PC1, the succulent/drought resistant taxa *Pentzia*-type and *Euphorbia*, as well as *Stoebe*-type, are all associated with high negative loadings. While the only taxon with a significant positive loadings is Myrtaceae which is a similar outcome to that found in the first PCA.

Both Ericaceae and Celastraceae show significant positive loadings for PC2 while a significant negative loading is seen for the coastal thicket element *Morella*.

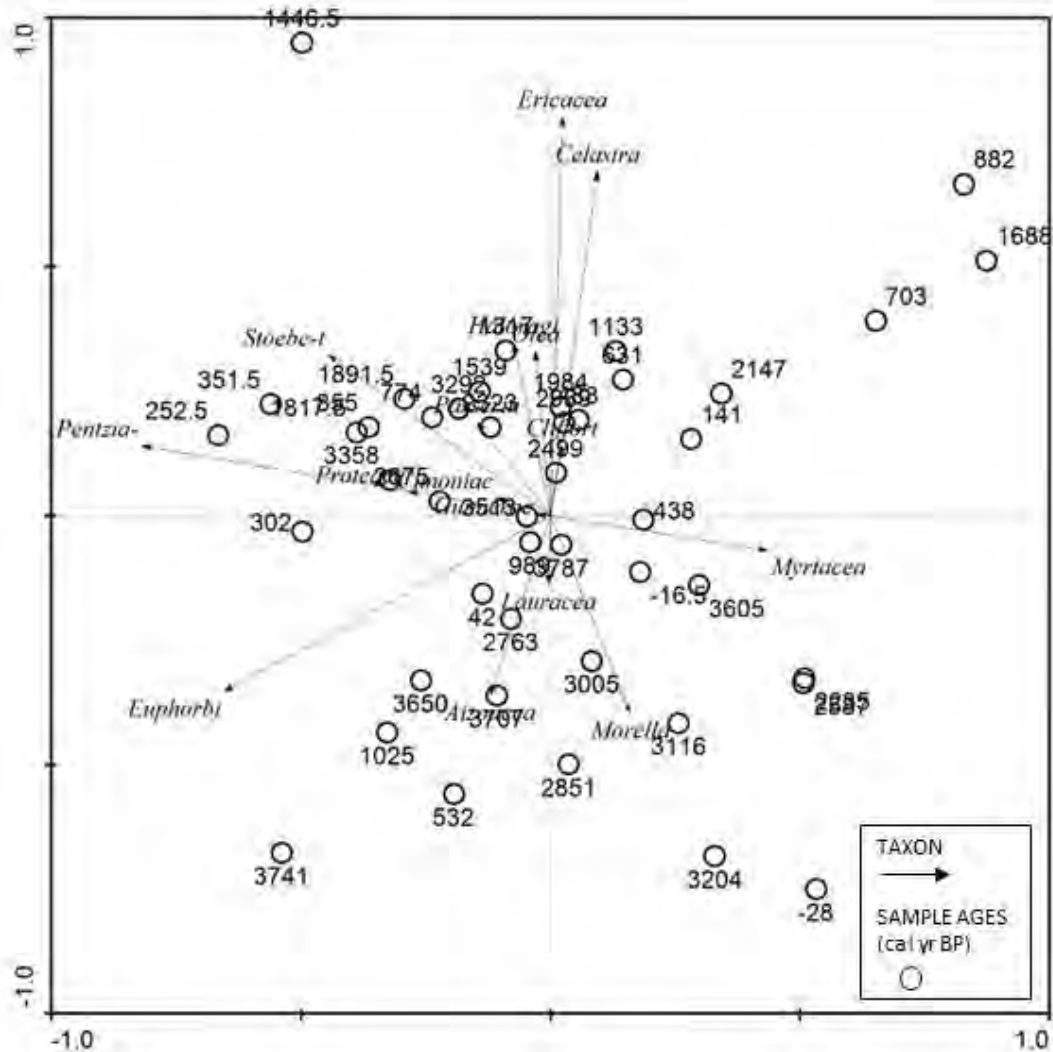


Figure 5.9. Ordination biplot for the two principal components from PCA 2, with PC1 on the x-axis and PC2 on the y-axis

As with the component loadings in Table 5.5, Figure 5.11 shows *Pentzia*-type and Myrtaceae on opposite sides of the plot. The same can be seen for Ericaceae, Celastraceae and Morella – indicative of the significant loadings for PC2. It is further noted that all the fynbos components are found in the top half of the plot, all in the upper left hand quadrant with the exception of Ericaceae. The three succulent/drought resistant components are all situated on the left side of the plot, with all the coastal thicket elements, except for *Olea*, on the right.

5.6. Conclusion

This chapter presented the palaeoecological results for EV1.11. From the radiocarbon results the EV1.11 record encompasses the period c. 3790 cal yr BP to 1980 AD – the late Holocene. The pollen results further clearly reflect the dominance of fynbos and succulent/drought resistant taxa in the assemblage. These results are discussed in more detail in the following chapter.

6. Discussion and Palaeoenvironmental Synthesis

6.1. Introduction

The EV1.11 record spans the late Holocene, marked by several events of “rapid climate change” (RCC) including the neoglacial, the Medieval Climate Anomaly and the Little Ice Age (Mayewski et al. 2004).

Chapter 6 discusses the results in terms of the climatic and environmental significance of major indicator taxa to provide a picture of late Holocene changes in the Wilderness region.

The chapter begins with an overview of the EV1.11 record, followed by a ~4000 year palaeoenvironmental reconstruction. Subsequently, patterns and trends identified in the EV1.11 record, specifically related to the above RCC events, are explored further. Finally, the EV1.11 record is discussed in relation to records from the three major rainfall zones of South Africa.

6.2. General Overview of Pollen Assemblage

The main contributors to the EV1.11 assemblage are the fynbos and succulent/drought resistant (SD) groups. Ericaceae, Restionaceae and *Stoebe*-type form the largest part of the fynbos group, with lesser contributions from *Passerina* and *Cliffortia*.

The large SD group sums are due to the high *ChenoAm*-type levels – here assumed to be more of an indicator of marine/saltwater conditions than drought as such. Thus the SD group can also be evaluated without the *ChenoAm*-type component which may possibly obscure other signals present in the data. Cyperaceae makes up the largest part of the aquatics/riparian vegetation group. Being a part of the local wetland vegetation, Cyperaceae is often overrepresented in the pollen record and as such its significance in relation to climatic and environmental inferences can be overestimated (Moore et al. 1991; Jackson & Williams 2004). The aquatic group thus, as with the SD group, can also be evaluated without Cyperaceae.

The smaller contributions to the assemblage from the Afrotropical forest (AF) group are mainly due to the pollen dispersal methods of the varying taxa as well as the low pollen production of forest taxa – except of course for *Podocarpus* which, due to its grain morphology, is very successfully distributed through wind (Jackson & Williams 2004). As a result, *Podocarpus* is also the largest component of the forest group sum, the forest group is then also further investigated with the *Podocarpus* contribution removed. The coastal thicket group also makes up a smaller part of the assemblage, with the main constituents of the group being *Olea* and *Rhus*.

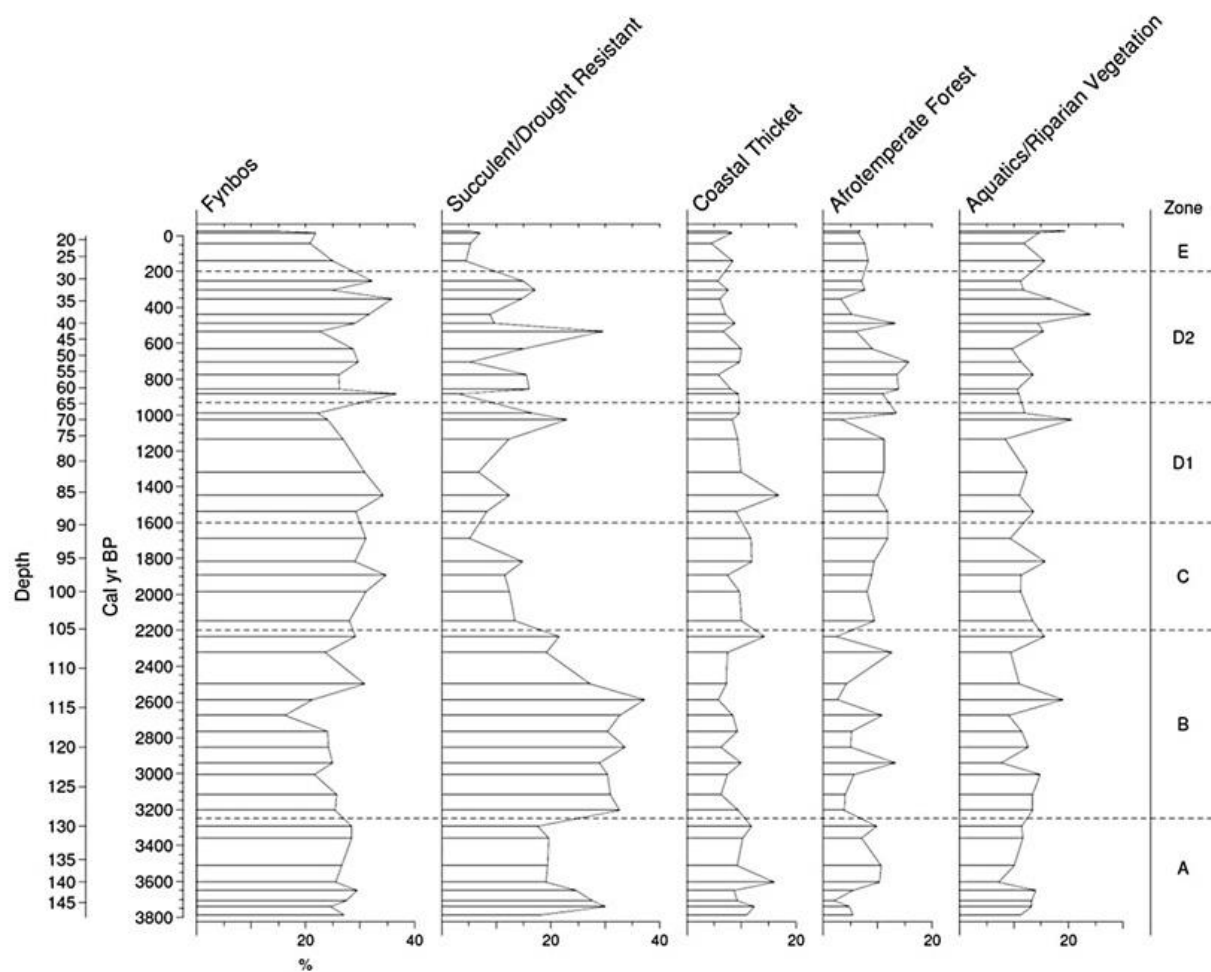


Figure 6.1. The ecological affinity group sums for EV1.11

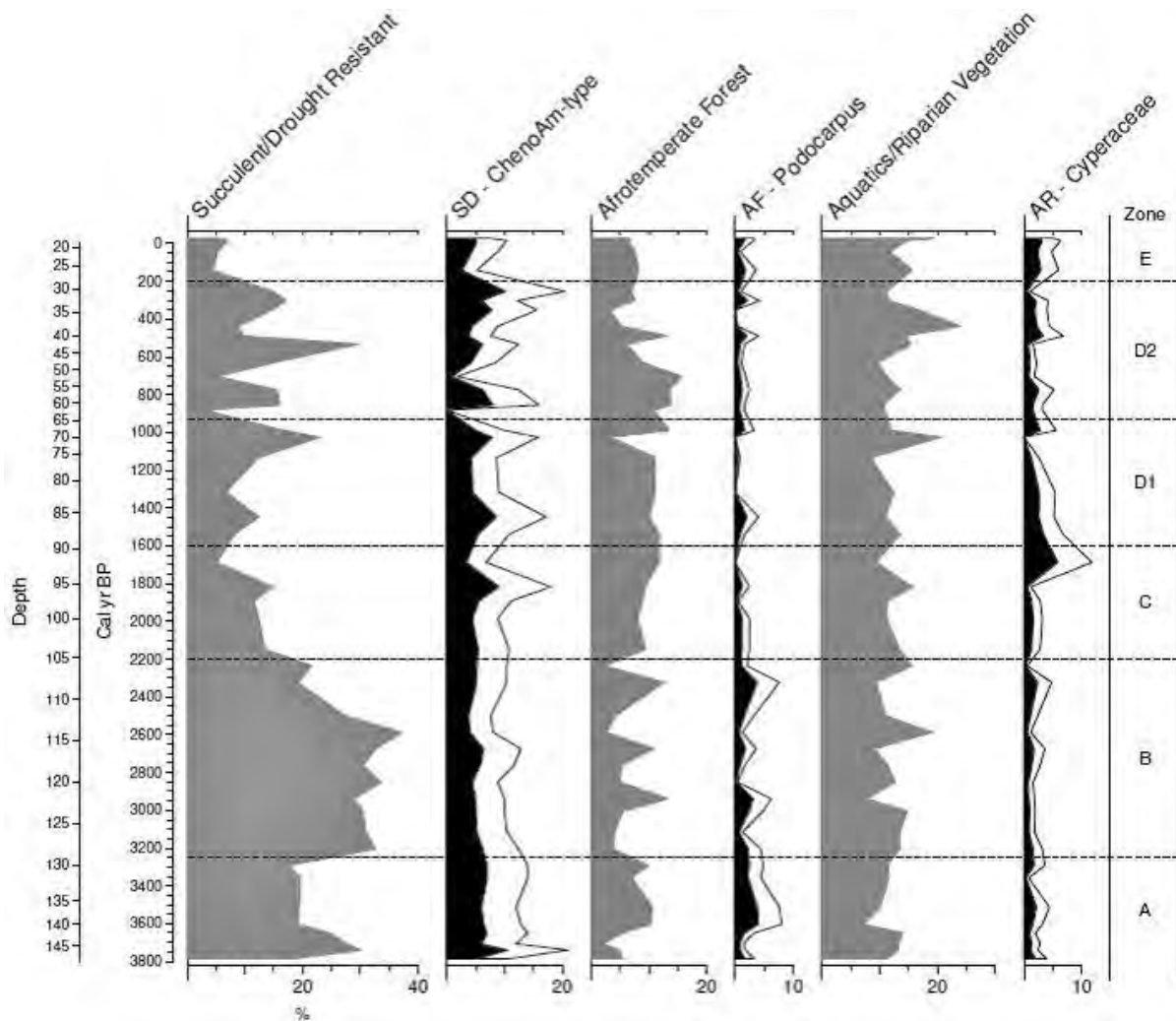


Figure 6.2. The Succulent/Drought resistant, Afrotropical Forest and Aquatics groups, with the largest contributing taxon to each removed from the total

6.3. Fire-Vegetation Dynamics

The literature on fire-vegetation dynamics (e.g. Cowling 1992; Bond 2004; Midgley et al. 2004) reaffirms the important relationships between fynbos and fire, with fynbos community development dependent on regular fire events.

Fire plays a further important role in maintaining fynbos-thicket and fynbos-forest boundaries. When fires are absent for long periods of time, thicket taxa can encroach on fynbos-dominated sites and progressively replace fynbos in the area (Privett & Lutzeyer 2010). The same is evident for the fynbos-forest relationship with forest species replacing surrounding fynbos when fire is excluded from the system (Midgley et al. 2004).

The charcoal record (Figure 5.9) shows relatively large fluctuations with several significant peaks in the charcoal concentration - most notably at c. 3600, c. 3300 and c. 2900 cal yr BP. These peaks,

which most probably represent large fire events, could be due to lightning and storm conditions as lightning-initiated fires are very common in this area (Horne 1981). As mentioned in section 2.5.2, bergwinds, and the associated dry conditions, are also significant driving forces relating to fires in the southern Cape (Geldenhuys 1994). With bergwinds most prevalent during winter, it may instead have been cool and dry with bergwinds responsible for these large fire events.

This being said, the 'Great fire of 1869' – a bergwind driven fire - is not depicted in the fire record from EV1.11. This may be due to the sampling resolution with the ages of the two samples encompassing this period interpolated to be 1809 and 1908 AD, respectively. Anthropogenic disturbances may also further play a role with regard to the deposition of charcoal fragments as European colonisation in this area was already well under way. In relation to this, Clark (1988) does point out that individual fire events are not always fully revealed in fossil charcoal records.

Figure 6.2 displays the fire-fynbos interactions in EV1.11. The fynbos and charcoal curves (A and B) do covary through most of the sequence. Notable exceptions are seen from c. 3740 to 3360 cal yr BP, 2850 to 2320 cal yr BP and again at c. 350 to 300 cal yr BP where increased charcoal concentrations coincide with lower fynbos values and *vice versa*. It is proposed that climate, for example decreased moisture availability and/or warmer temperatures, may have been the limiting factor during these periods, prohibiting sufficient establishment of the relevant fynbos communities.

Ericaceae (C), being a "true" fynbos component, shows a more direct positive relationship with charcoal. It is further noted that similar fluctuations are observed in the Ericaceae and charcoal series, specifically the periods highlighted above where fynbos and charcoal seem to be out of phase. At c. 2150 cal yr BP. However, the largest Ericaceae values within the assemblages coincides with lower charcoal values. As mentioned above, changes in moisture availability and/or temperature may have been the predominant factor governing vegetation change at this period. However, fynbos levels show a similar increasing trend as charcoal around this period.

The discord observed between the charcoal concentration and fynbos may further be due to the diversity of the individual taxa contained in this group. For example, *Passerina* can be classified as dune fynbos or dune shrub (Cowling 1983), and may respond differently to the effect of fire on "true" fynbos. This is further illustrated in Figure 6.2 where *Passerina* (D) displays an inverse relationship with the charcoal concentration. Exceptions are noted at c. 880 cal yr BP, where maximum proportions of *Passerina* coincides with moderately increased charcoal levels, as well as towards the top of the sequence, from c. 350 cal yr BP, where indications are that anthropogenic factors, rather than climate is responsible for changes in vegetation.

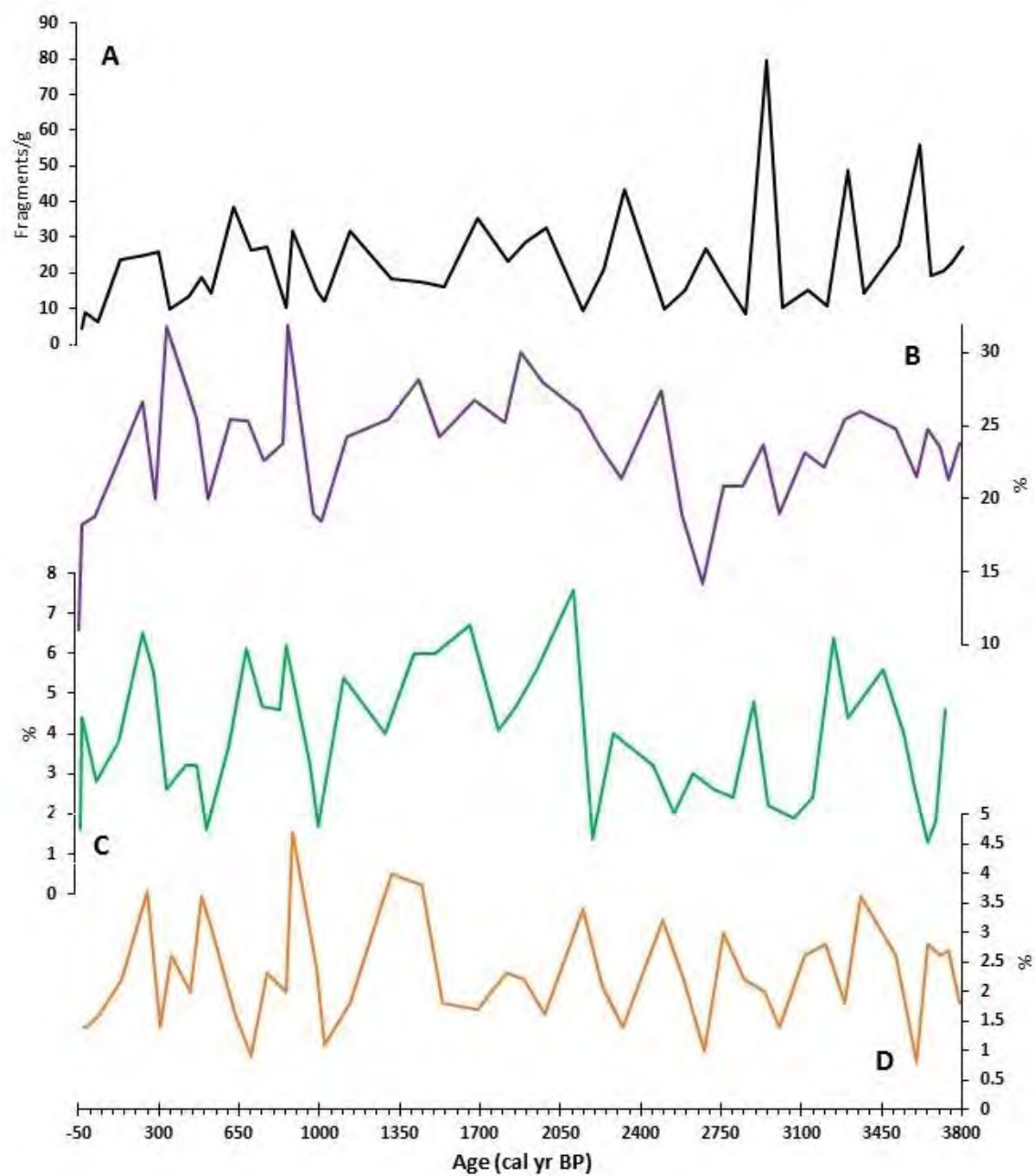


Figure 6.3. The relationship between fire and fynbos, A: charcoal concentration, B: fynbos group sum, C: Ericaceae, D: *Passerina*

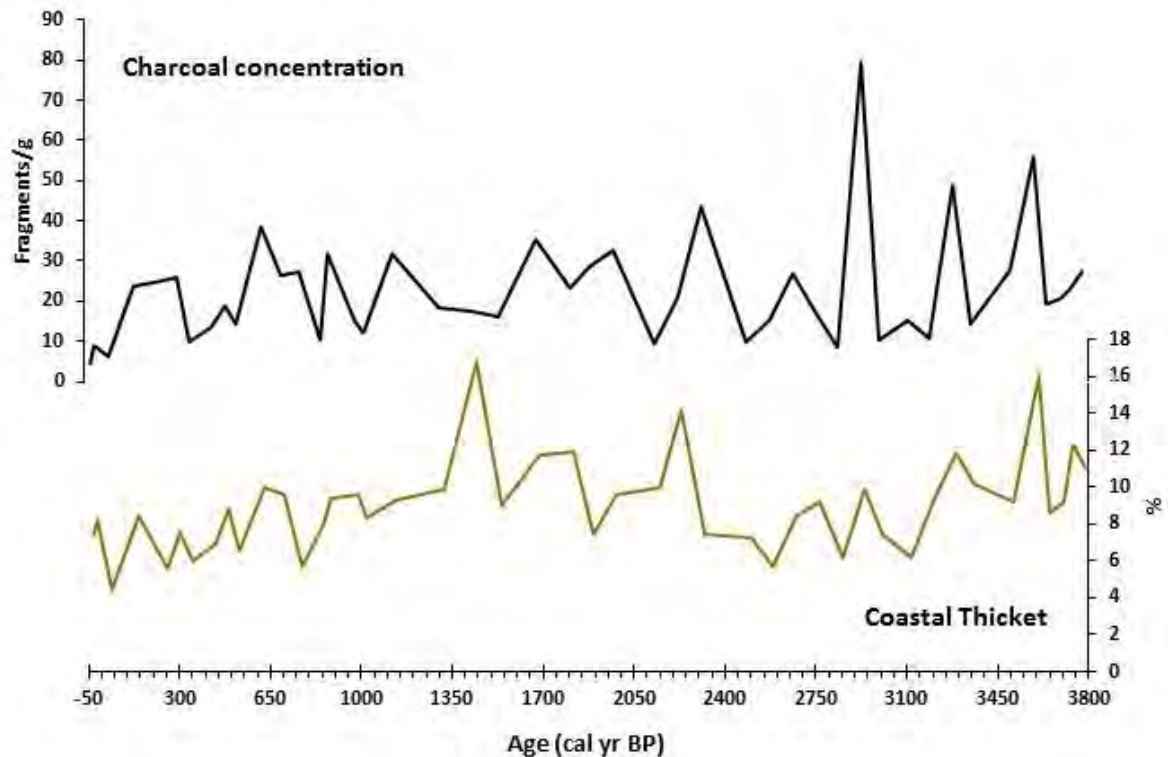


Figure 6.4. The relationship between fire and coastal thicket

In Figure 6.4 the relationship between fire and coastal thicket elements is demonstrated. Charcoal concentration is expected to have an inverse relationship with thicket taxa, and indeed this is true for certain periods within the record, e.g. a very large spike in thicket at c. 1450 cal yr BP with coincident lower values for charcoal. Around c. 2235 cal yr BP, thicket seems to show a lagged response to fire with low charcoal values at c. 2150 cal yr BP. There are also, however, periods with increases in both charcoal and thicket, most notably from the beginning of the record, c. 3790 cal yr BP, to around c. 2675 cal yr BP. This might be due to the influence of shifting dune sand on thicket taxa as this falls within the period identified by Martin (1968) when dunes were most mobile.

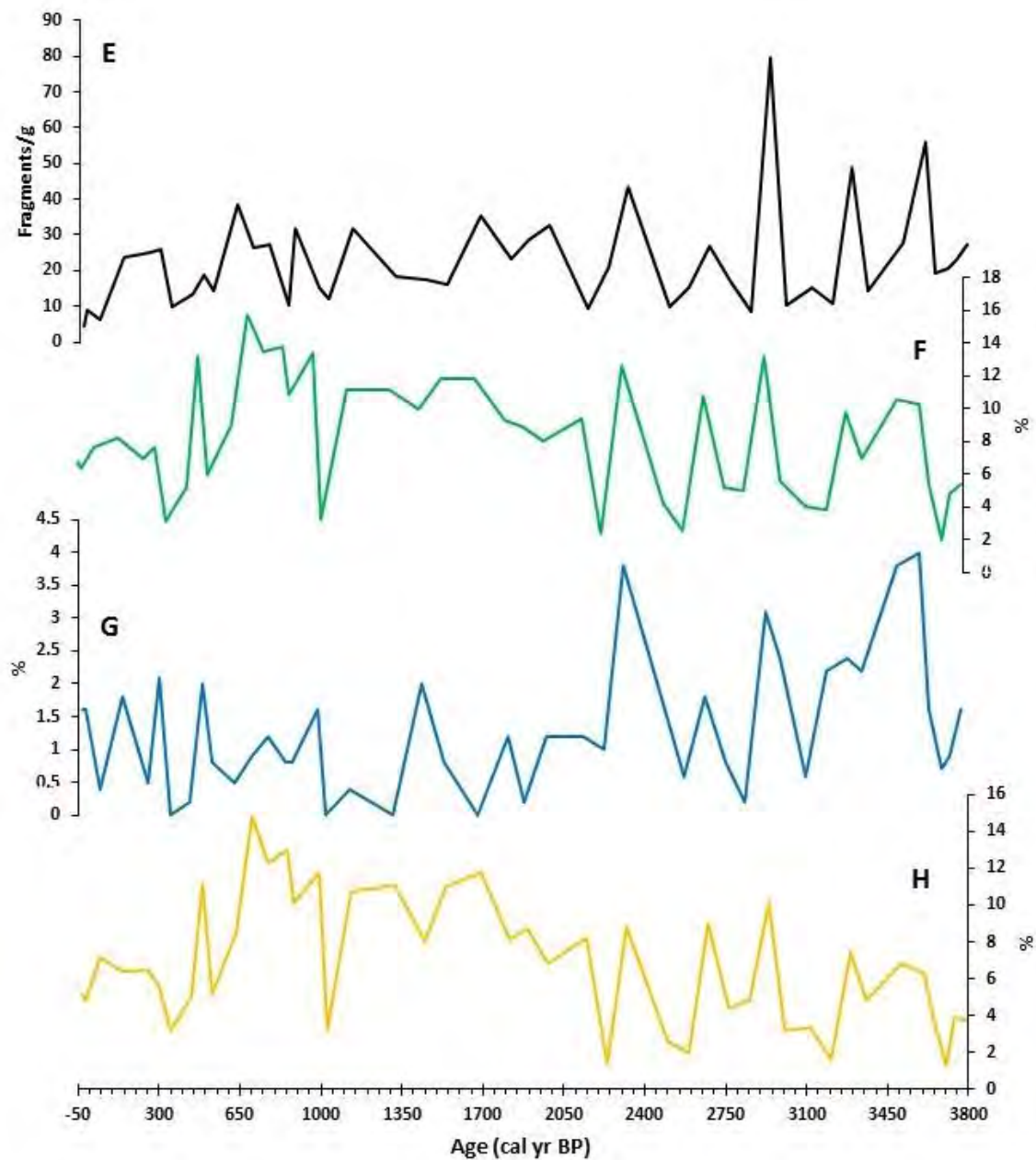


Figure 6.5. The relationship between fire and Afrotemperate forest, E: charcoal concentration, F: Afrotemperate forest group sum, G: AF – *Podocarpus*, H: *Podocarpus*

Plotting the Afrotemperate Forest group sum (F) against the charcoal concentration (E) (Figure 6.5), an unexpected positive relationship is seen for most of the sequence. When removing *Podocarpus* from the forest group sum (G), the anticipated inverse relationship between charcoal concentration and forest is displayed - most notably the period c. 2265 cal yr BP to 14 cal yr BP, or 1810 AD. It is interesting to note that, from the start of the sequence to c. 2235 cal yr BP, concurrent peaks and troughs in charcoal and forest is observed. This is also true for *Podocarpus* (H) and the forest group

sum with *Podocarpus* removed. This might again suggest that other factors, i.e. dune activity and/or changes in sea level, are at play governing vegetation dynamics.

An important detail to highlight is that at c. 2940 cal yr BP, the highest charcoal concentration in the sequence, peaks are observed for the ecological groups fynbos, coastal thicket and Afrotemperate forest. It is further noted that a general synchronicity is displayed between forest, thicket and charcoal for the start of the sequence up to around c. 2320 cal yr BP. This falls within the period where SD levels, mostly due to *ChenoAm*-type, are at its highest and considerable changes with regard to sea level and dune morphology were underway.

The records presented here do broadly agree with the suggested fire-vegetation dynamics but also highlight the complex interactions that are inherent to such dynamic systems.

6.4. Late Holocene Palaeoenvironmental Reconstruction for Eilandvlei

6.4.1. Temperature and Moisture Indices

On closer inspection of the PCA results, PC1 of the second PCA (with *ChenoAm*-type and *Podocarpus* removed from the data set) seems to be linked with moisture availability with high negative loadings for the drought tolerant taxa *Pentzia*-type and *Euphorbia* being most prominent. This suggests that high positive scores for PC1 indicates increased moisture availability while high negative scores indicate less moisture and possibly increased aridity. The second component, PC2, of PCA 2 shows a possible relation to relative temperature with high positive loadings associated with cold tolerant taxa such as Ericaceae and negative loadings, though not as high, shown for warm taxa for instance *Morella*, Aizoaceae and *Euphorbia*. High positive scores for PC2 thus indicate lower temperatures whereas higher negative scores point to increased temperatures.

PC1 and PC2 may accordingly be used as indices for moisture availability and relative temperature, respectively, in the reconstruction of the late Holocene at Eilandvlei, as set out below. It has to be kept in mind, however, that PCA results may have limited value, specifically with regard to pollen analysis, as patterns reflected in the pollen data may not always be explicitly expressed in the principal components (Quick, pers comm).

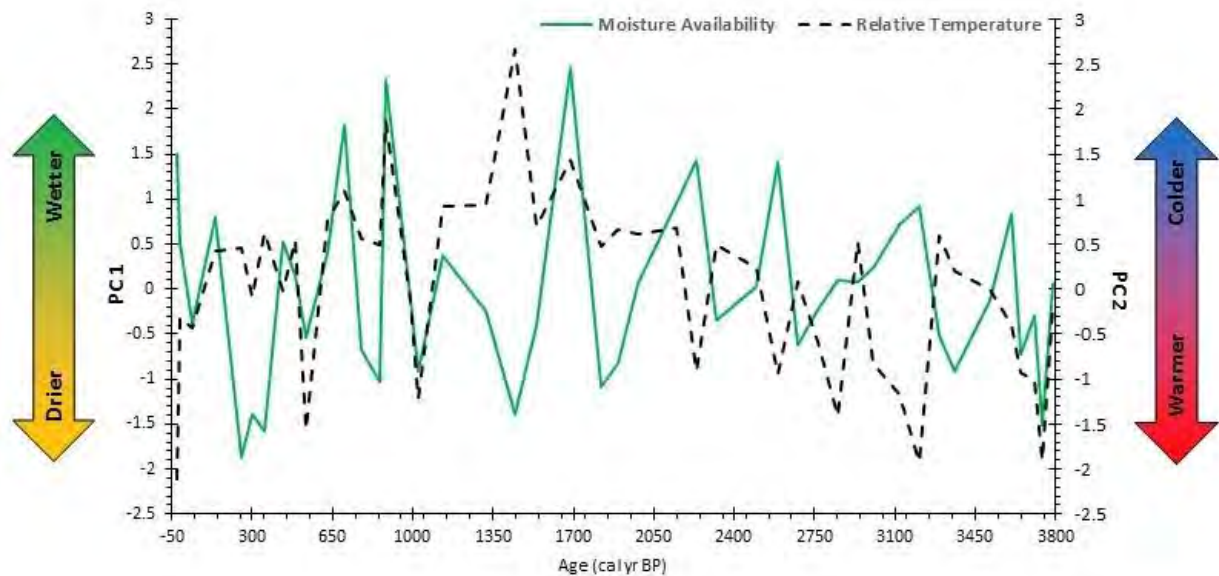


Figure 6.6. The moisture availability and temperature indices for EV1.11 as derived from Principal Component Analysis

6.4.2. c. 4000 years of Vegetation and Climate Change at Eilandvlei

Around the start of the record, c. 3740 cal yr BP, both the moisture and temperature indices point to a warm and dry environment. This is substantiated by the pollen record, with Aizoaceae displaying its maximum value within in the assemblage here with a further notable peak in the SD group. Indications are thus that the sequence commences with an arid event.

Following this dry episode, a substantial increase in moisture availability is indicated from the moisture index (PC1), most notably at c. 3600 cal yr BP, coincident with maximum values for the moist forest elements Moraceae and Myrsine, as well as the monolet spore group. *Podocarpus* is also more abundant for the 100 year period around this peak in moisture with the forest group as a whole showing a sharp increase from c. 3700 to 3500 cal yr BP. As *Podocarpus* is drought intolerant and as such is an indicator of increased moisture availability/wetter conditions, it reaffirms inferences from the moisture index. The Eilandvlei diatom record further confirms this with a markedly wetter period indicated from c. 3750 to 3650 cal yr BP (Kirsten 2014).

Increases are similarly observed for Ericaceae, Cyperaceae and Poaceae during this period and extending to c. 3300 cal yr BP, with Fynbos constantly present in high proportions. Onagraceae, a wet forest component, present at c. 3300 cal yr BP for the only time within the sequence, further confirms wetter conditions. From the abundant Fynbos presence it can be postulated that an increasing winter rainfall influence prevailed in the Wilderness region until c. 3300 cal yr BP. Charcoal concentration further displays concomitant peaks with Ericaceae at c. 3600 and 3300 cal yr BP, indicating fire events

during this generally wetter period. These fires might be the result of bergwinds, most prevalent during winter and a major cause of fire in the region (Geldenhuys 1994). Scholtz (1986) does however note increased summer rainfall from c. 4400 to 2600 cal yr BP in this region.

A rapid decrease in coastal thicket is seen from c. 3600 to 3500 cal yr BP, followed by a marginal recovery c. 3300 cal yr BP, from where the declining trend continues to around c. 3100 cal yr BP. This pattern is emulated in the forest group with a similar decline in aquatic taxa from c. 3500 to 3360 cal yr BP. Kirsten (2014) notes a short-term dry episode and resultant lower lake levels at Eilandvlei c. 3650 to 3550 cal yr BP which might explain the above evidence. The SD group however shows no marked response during this period with only a slight increase observed from c. 3500 to 3360 cal yr BP.

Significant variability is further observed in the pollen record during this period, most notably for the forest group between c. 3100 and 2150 cal yr BP, and coastal thicket from c. 3100 to 2300 cal yr BP. The fynbos group exhibits similar fluctuations although the magnitude of change is not as great. The trends, or lack thereof, pointed out above can most likely be attributed to extensive dune movement as this time falls within the period of increased dune activity in the region - c. 3700 to 2400 cal yr BP – as identified by Bateman et al. (2011). Martin (1968) furthermore suggests that this period of greater dune activity extended to around c. 1900 cal yr BP. Hence, as proposed by Martin (1968), the observed vegetation changes can rather be ascribed to large scale dune movements than as a response to a drier and/or windier environment.

Around the same time, c. 3200 to 2500 cal yr BP, *ChenoAm*-type dominates the sequence. This is most likely the expression of a marine incursion at Eilandvlei where Martin (1968) identified a significant marine influence at Groenvlei from c. 7000 to c. 3200 cal yr BP. Kirsten (2014) similarly recognised a marine transgression at Eilandvlei until c. 3750 cal yr BP, with the maximum expression of this event placed at c. 3900 cal yr BP. This marine influence further makes brief reappearances around c. 3425 and 2900 cal yr BP (Kirsten 2014). The constant presence of both Poaceae (possibly *Phragmites*) and Juncaceae together with *ChenoAm*-type during this period can additionally suggest a more saline environment (Baxter & Meadows 1999).

With the overriding *ChenoAm*-type presence as well as the effect of aforementioned dune mobility, other signals within the assemblage may possibly be obscured. It can however be inferred that the period c. 3500 to 2100 cal yr BP represents a highly dynamic environment.

A substantial decline in *ChenoAm*-type pollen is observed from c. 2600 to around c. 1700 cal yr BP. Around c. 2200 cal yr BP, a slight peak in *ChenoAm*-type proportions are however displayed, breaking

the overall declining trend. Coincident with this, a notable drop in both *Podocarpus* and the forest group as a whole, as well as aquatic elements are seen. Ericaceae is likewise present in very low proportions with a significant peak in *Anthospermum*-type further observed. Kirsten (2014) proposes drier conditions at Eilandvlei prior to c. 2100 cal yr BP, with a warmer environment from around c. 2375 cal yr BP. The temperature index similarly points to warmer conditions at c. 2200 cal yr BP further supporting a period of increased aridity, albeit brief.

Fynbos shows a general increasing trend from around c. 2670 to 1900 cal yr BP with a notable peak at c. 2500 cal yr BP – concomitant with the onset of the *ChenoAm*-type decline. A prompt, sharp increase in Ericaceae is observed from c. 2200 to 2150 cal yr BP, displaying its highest value within the assemblage here. Ericaceae remains present in high proportions until c. 1100 cal yr BP, with only marginally lower levels noted at c. 1800 and 1300 cal yr BP. The above evidence can suggest cooler, and potentially wetter, conditions and thus a possible shift towards a winter rainfall regime. The presence of *Oxalis* from c. 1980 to 1800 cal yr BP might further substantiate this.

The increased presence of *Typha* during this period, coupled with lower *ChenoAm*-type levels, can further indicate fresh water input into the system. Indications are thus that the influence from the marine transgression is starting to dissipate with a return to fresh water conditions. In this regard, Kirsten (2014) defined the period c. 2100 to 2000 cal yr BP as a transitional phase with Eilandvlei changing from a lagoon to coastal lake, with a similar transitional stage present at Groenvlei (Martin 1968).

A major peak in *Typha* is seen at c. 1700 cal yr BP. Poaceae, Ericaceae and *Olea* are further present at notably high levels. The temperature and moisture indices additionally point to colder and wetter conditions than earlier. *Podocarpus* is abundant at this point, although no other forest elements are present here. A major wet phase can accordingly be inferred at c. 1700 cal yr BP, possibly the maximum expression of winter rainfall in the region. Kirsten (2014) similarly mentions a wetter environment after the transitional phase c. 2000 cal yr BP, persisting until around c. 1700 cal yr BP.

Increased *Podocarpus* levels are seen from c. 2200 cal yr BP to around c. 1100 cal yr BP, with sporadic appearances of other wet forest elements, i.e. *Cunonia*, Icacinaceae and *Ilex* during this period. This coincides with Martin's (1968) "last phase of forest expansion" which extended from around c. 1900 to 900 cal yr BP.

In contrast to indications of forest expansion and a wetter environment at Eilandvlei and Groenvlei, the Norga record shows a decline in forest elements from c. 2600 to c. 1300 cal yr BP, with Scholtz (1986) proposing a decrease in summer rainfall resulting in a more arid environment.

Large peaks in *Pentzia*-type, *Crassula* and *Anthospermum*-type, indicators of dry and/or warm conditions, are observed at c. 1450 cal yr BP in the EV1.11 record. The fynbos group registers a concurrent decline from c. 1500 to c. 990 cal yr BP with *ChenoAm*-type levels steadily on the rise from c. 1300 cal yr BP, peaking at c. 1000 cal yr BP. A marked decline is further observed for the aquatics group, after the wet event at c. 1700, to c. 1000 cal yr BP. Increased dune activity is also again noted in the region from c. 1750 to 1300 cal yr BP (Bateman et al. 2011) which, in conjunction with the diatom evidence, lead Kirsten (2014) to propose cooler and/or windier, and very possibly drier, conditions in the area. Accordingly it can be posited that the arid period, as identified at Norga, started later and finished earlier in the Wilderness region.

From c. 1100 to 1000 cal yr BP, a notable decrease in *Podocarpus* and the forest group in general is observed, with *Podocarpus* the only forest element present at c. 1000 cal yr BP. Ericaceae further displays a marked decrease with the SD group registering a large peak. *ChenoAm*-type similarly peaks after its decreased presence from c. 2600 to 1700 cal yr BP.

This can indicate a dry episode with the PCA further pointing to warmer and possibly drier conditions, although short-lived, as a prompt increase in forest taxa are observed with levels peaking at c. 990 cal yr BP. The aquatics group likewise displays a peak at c. 990 cal yr BP and remains present at relatively high levels until c. 700 cal yr BP. *Podocarpus* levels further increase to reach maximum proportions at c. 700 cal yr BP, with only a very marginal drop noted at c. 880 cal yr BP. The other forest elements do however exhibit a generally decreasing trend until c. 630 cal yr BP.

Fynbos similarly shows a rapid increase from c. 1000 cal yr BP, reaching maximum levels at c. 880 cal yr BP. The SD group simultaneously achieve minimum values and *ChenoAm*-type shows a further notable decline. The SD group does however peak again at c. 850 after being completely absent from the assemblage at c. 880 cal yr BP, with another swift decline to c. 700 cal yr BP.

As is evident above, the period c. 1100 to 700 cal yr BP is marked by considerable variability in the pollen record, suggesting unstable climatic conditions. The timing of this variable period coincides with the globally recognised Medieval Climate Anomaly (MCA) – identified in South Africa to encompass the interval c. 1050 to 650 cal yr BP (Tyson & Lindesay 1992; Tyson 1999a, b; Tyson & Preston-Whyte 2000; Mayewski et al. 2004) - which could be responsible for the observed climatic and environmental instability.

Kirsten (2014) does however place the onset of the MCA in the region at around c. 1200 cal yr BP, while further suggesting increased aridity in the region from c. 1000 to 820 cal yr BP. The Groenvlei

pollen record further registers a decrease in arboreal taxa around c. 850 cal yr BP (Martin 1968) which is not explicitly evident in the EV1.11 pollen record.

After the peak in fynbos, c. 880 cal yr BP, levels decrease until c. 530 cal yr BP - a minor recovery is noted between c. 770 and 630 cal yr BP. Ericaceae shows a similar decline from c. 700 to 530 cal yr BP, after its ample presence from around c. 990 to 700 cal yr BP. Around c. 530 cal yr BP, low levels for the forest group marks the end of the rapid decline observed after the maximum at c. 700 cal yr BP. The increased occurrence of succulent and drought resistant taxa are further noted at c. 530 cal yr BP with *ChenoAm*-type similarly present in high proportions. This major peak - the highest proportion of *ChenoAm*-type since the marine transgression c. 3200 to 2500 cal yr BP - can indicate a local sea level effect and a possible return to estuarine conditions. Kirsten (2014) indeed notes a possible increased marine influence in Eilandvlei c. 600 to 500 cal yr BP which supports this interpretation. In addition to the greater marine influence, the evidence above suggests warmer and most likely drier conditions with lower lake levels further reported at both Eilandvlei and Swartvlei c. 500 to 300 cal yr BP (Kirsten 2014).

Even though it seems that the onset of the Little Ice Age (LIA) is not specifically reflected in the pollen record, the drier period prior to c. 530 cal yr BP is most probably the manifestation of the warm event separating the two colder phases of the LIA. This being said, it is interesting to note that both *Anthospermum*-type and *Podocarpus* reach maximum values within the assemblage at c. 700 cal yr BP. *Anthospermum*-type, typically being an indicator of drier conditions, and *Podocarpus* generally an indicator of increased moisture, provide a conflicting picture. This peak in *Anthospermum*-type accompanied by the increased presence of Ericaceae and relatively high levels of the fynbos group, can however imply a colder, more winter rainfall dominated environment, possibly marking the onset of the LIA in the region.

Following this drier episode, an abrupt increase in moisture availability is demonstrated in the record. *Podocarpus*, and other wet forest elements, i.e. *Cunonia*, *Ilex*, reach markedly high levels at c. 490 cal yr BP, with a concurrent peak observed for the aquatics group. Cyperaceae, Juncaceae and Restionaceae furthermore reach maximum proportions within the sequence at c. 440 cal yr BP. Ericaceae levels display an increasing trend from c. 530 to 250 cal yr BP with *Oxalis* persistently present between c. 530 and 350 cal yr BP. The fynbos group similarly increases to reach near maximum proportions at c. 350 cal yr BP. A major peak in *Typha* is further observed at c. 300 cal yr BP with the aquatics group present at higher levels but generally decreasing until c. 250 cal yr BP. Accordingly it can be inferred that this increased moisture availability is possibly the result of increased winter rainfall in the region.

Even though the evidence points to increased moisture availability in the region, both *Podocarpus* and the forest group registers a rather rapid decline after peaking at c. 490 cal yr BP, culminating at c. 350 cal yr BP with *Podocarpus* the only forest element present.

From approximately c. 500 cal yr BP, Reinwarth et al. (2013) notes an increasing trend in the concentration of fine minerogenic sediment components in Eilandvlei with a resurgence in dune activity in the Sedgefield region from c. 600 to 430 cal yr BP (Bateman et al. 2011). It can therefore be postulated that the extensive sand movement associated with increased dune activity, can possibly serve to explain the inconsistencies observed in the EV1.11 record between c. 500 and 350 cal yr BP. In addition, *Stoebe*-type, a classic disturbance indicator in fynbos (Meadows et al. 1996), is present at maximum levels at c. 350 cal yr BP, further lending credence to the idea of large scale environmental disruptions.

The fynbos group exhibits a steady decline after the peak at c. 350 cal yr BP with Asteraceae HS present in high proportions (near the sequence maximum) between c. 300 to 250 cal yr BP (1650 to 1700 AD). *ChenoAm*-type further displays increased values around c. 300 cal yr BP with the rest of the SD group increasing to reach near maximum values at c. 250 cal yr BP.

In association with the evidence for increased moisture availability from c. 530 to around 350 cal yr BP, it may be inferred that the second phase of the LIA was initiated with a cold and wet episode, becoming progressively drier towards c. 250 cal yr BP – this point may possibly represent the termination of the LIA in the area. A peak in forest taxa are however again observed at c. 300 cal yr BP, after being completely absent from the assemblage at c. 350 cal yr BP. This may again be a remnant of dune activity in the region.

Towards the top of the record, inferences based on vegetation changes become more tentative as signals can be obscured by the influence of anthropogenic factors and thus climate alone cannot be considered the driving force behind the observed changes.

The onset of European colonisation in the EV1.11 record is marked by the appearance of *Pinus* at c. 140 cal yr BP (1810 AD). Historical reports date the arrival of European settlers in the region to the 1700's (Phillips 1931; Van Daalen 1980). This discrepancy with the EV1.11 record is most probably an artefact of the sampling resolution - the interpolated ages from the age-depth model indicate that there are just over 100 years between samples, with the distance between samples ± 3.5 cm. This period is further marked by an exceptionally higher deposition rate compared to the rest of the core, most certainly due to increased erosion brought about by enhanced agricultural and logging activities.

Forest group proportions, specifically *Podocarpus*, remain relatively constant since the introduction of *Pinus* to the environment, although minor fluctuations in the other components of the group are seen. Here again, the sampling resolution has to be kept in mind as only four samples represent this period and thus signals within the record may be obscured.

Another result of the increased human presence identified in the EV1.11 record is the substantial decline of Poaceae from c. 140 cal yr BP (1810 AD) pointing to the introduction of livestock and the associated degeneration of grassland due to overgrazing. Numerous drought years have however been reported for the late 1800's in this area (Phillips 1931; Van Daalen 1980) although this arid period is not clearly reflected in the EV1.11 record. Nonetheless, there are some indications of this such as increased *Anthospermum*-type, a slight increase in SD elements, most notably *Euphorbia*, a general decline in fynbos as well as the above mentioned decrease in Poaceae. This may of course rather be a reflection of the anthropogenic influence on the environment with *Euphorbia*, a non-palatable karroid succulent, progressively replacing grassland and fynbos (Meadows et al. 1996; Meadows & Baxter 2001) in appropriate localities

6.5. Marine Influence versus Moisture Availability

From the principal component analysis (section 5.5) a very strong inverse relationship between *ChenoAm*-type and *Podocarpus* was alluded to. This is illustrated in Figure 6.7.

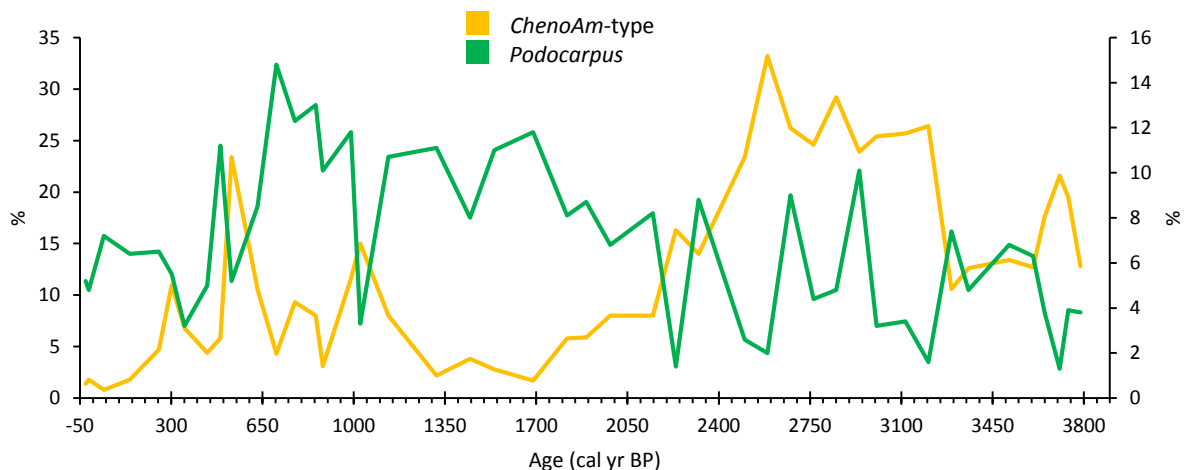


Figure 6.7. The inverse relationship between *ChenoAm*-type and *Podocarpus*

At first glance, the observed pattern is expected as *Podocarpus* is a moist forest component, indicative of increased moisture availability, while *ChenoAm*-type is usually taken to be indicative of drier

conditions (Scott 1999). Alternatively, especially along the coast, *ChenoAm*-type can suggest salt marsh or estuarine conditions (Neumann et al. 2011b) while further possibly being an indicator of increased marine conditions (Meadows & Baxter 2001).

Due to the inferred marine incursion at c. 3 200 cal yr BP and the location of Eilandvlei relative to the coast, the presence of *ChenoAm*-type pollen is here taken to be indicative of a trend towards an increased marine influence in the lake. But, this marine influence alone cannot be responsible for the contrasting curves in Figure 6.7.

The Seweweekspoort (SWP) $\delta^{15}\text{N}$ record provides a good indication of moisture variability in the Southern Cape during the Holocene, with enriched $\delta^{15}\text{N}$ associated with increased aridity (Chase et al. 2013). The Eilandvlei diatom record can likewise be used to infer moisture availability, where an increase in brackish taxa relates to increased moisture availability (Kirsten 2014). Subsequently, these records will be used to try and resolve the marine-moisture conundrum.

The *ChenoAm*-type and SWP $\delta^{15}\text{N}$ series largely exhibits a positive relationship (Figure 6.8), most notably from around c. 3200 to 2000 cal yr BP, and c. 1000 to 300 cal yr BP. For the period c. 1940 to around c. 970 cal yr BP however, both series deviate from this trend displaying an inverse relationship.

Negative correlations are observed between the *ChenoAm*-type and brackish diatom series (Kirsten 2014), the most pronounced periods are from c. 3240 to 1080 cal yr BP and again between c. 830 to 500 cal yr BP. Two short lived intervals where positive correlations are displayed are noted from c. 1080 to 940 cal yr BP and c. 530 to 410 cal yr BP.

A general negative correlation is seen for the *Podocarpus* and SWP $\delta^{15}\text{N}$ records. This is expected as depleted $\delta^{15}\text{N}$ is taken to be indicative of increasing humidity, or moisture availability, (Chase et al. 2013) with a resulting increased *Podocarpus* presence. Exceptions are however noted at c. 3300 to 3200 cal yr BP as well as for the period between c. 2150 and 1025 cal yr BP where similar trends are present in the records.

Positive correlations are observed between *Podocarpus* and the brackish diatom series. From c. 3000 to 2240 cal yr BP however, the *Podocarpus* record is marked by significant fluctuations which makes it difficult to identify a general trend in the series. As with *ChenoAm*-type, there are short periods where the records deviate, these are constrained to the intervals c. 3420 to 3250 cal yr BP and c. 440 to 250 cal yr BP.

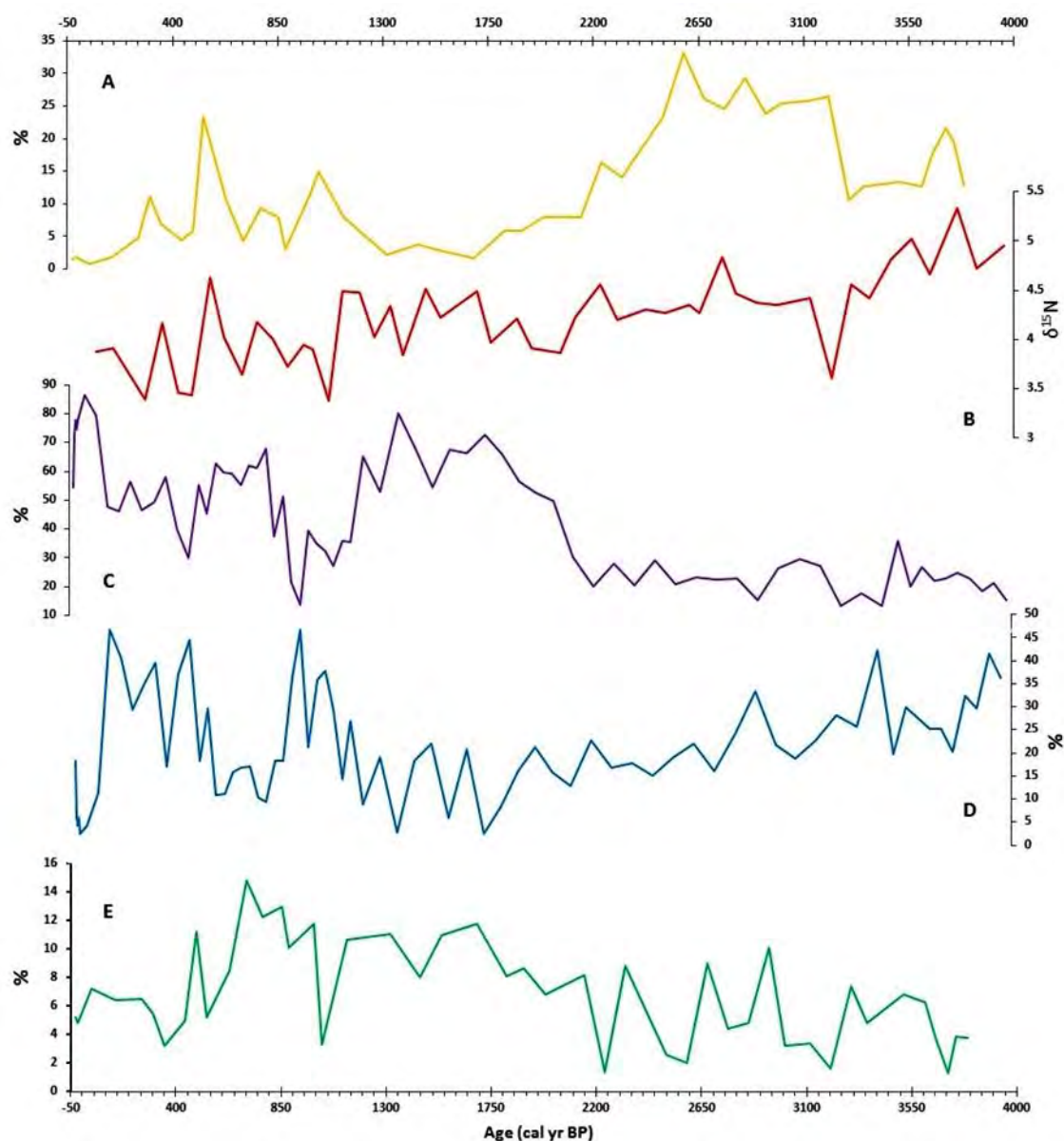


Figure 6.8. Composite graph exploring moisture availability in the Southern Cape, A: EV1.11 *ChenoAm*-type record, B: Seweweekspoort $\delta^{15}\text{N}$ record (Chase et al. 2013), C: Eilandvlei brackish diatom record (Kirsten 2014), D: Eilandvlei marine diatom record (Kirsten 2014), E: EV1.11 *Podocarpus* record

Looking at the marine diatom taxa and *ChenoAm*-type relationship some interesting patterns emerge. A rapid increase in *ChenoAm*-type levels are seen from the start of the record, c. 3790 cal yr BP, to c. 3700 cal yr BP. This peak is however only after a prompt decline in marine diatom proportions, from

maximum values at c. 3900 cal yr BP to the termination of the marine transgression c. 3750 cal yr BP, as identified by Kirsten (2014).

A plausible explanation may be the different response times of vegetation and diatom communities to changes in the environment and/or climate with diatoms being more sensitive, especially relating to local hydrology (Kirsten 2014). Consequently, these changes will be reflected in the diatom record prior to it being evident in the pollen record.

A large peak for the marine diatom group is displayed at c. 3425 cal yr BP followed by a steady decline to c. 3070 cal yr BP before peaking again at c. 2900 cal yr BP. Kirsten (2014) relates these two peaks to a brief resurgence of marine conditions in the lake. *ChenoAm*-type levels show a marked sudden increase from c. 3300 to 3200 cal yr BP. A further gradual increase is seen, culminating at c. 2600 cal yr BP where *ChenoAm*-type proportions are at a maximum. The rapid increase might be in response to the returning marine influence at c. 3425 cal yr BP. A similar argument can be made for the *ChenoAm*-type peak at c. 2850 cal yr BP, following the marine diatom peak at c. 2900 cal yr BP.

However, for the period between c. 3200 and 2600 cal yr BP when *ChenoAm*-type are at a maximum, the marine diatom group exhibits a general declining trend – with the exception of the peak at c. 2900 cal yr BP. Significant variability is present in the marine diatom record from around c. 2200 cal yr BP but looking at the broader trends, both series decline from c. 2190 to 1355 cal yr BP followed by and increase to c. 1040 cal yr BP. Correspondence between the two records are again noted from around c. 580 to 540 cal yr BP.

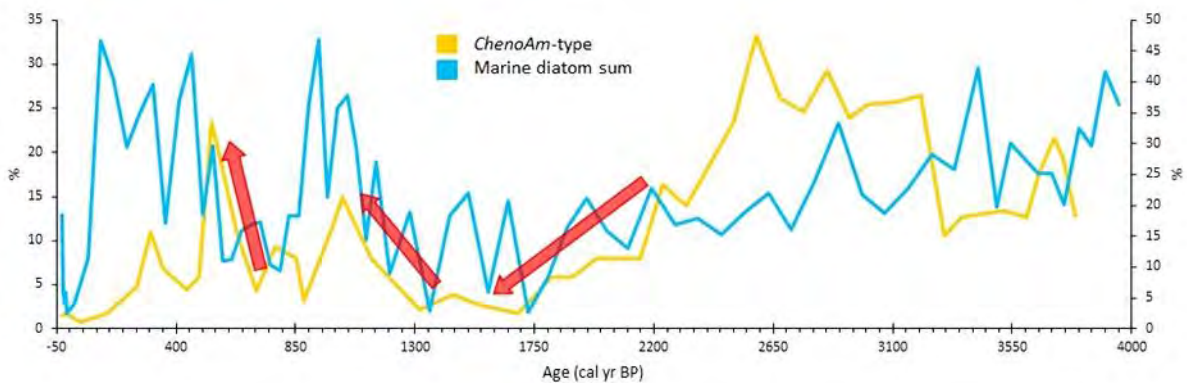


Figure 6.9. The general trends observed in the *ChenoAm*-type and marine diatom records from c. 2200 cal yr BP

Taking the time lags into account, the *ChenoAm*-type and marine diatom record do demonstrate rather strong agreement. However, with the deviation between the two series from c. 3200 to 2600 cal yr BP, as well as the timing of the marine transgression, as inferred in each record, differ by ~600 years, the marine influence on *ChenoAm*-type taxa was most likely over-estimated.

Furthermore, with the strong correlations between *ChenoAm*-type and both the SWP $\delta^{15}\text{N}$ and brackish diatom records, *ChenoAm*-type abundance would then rather be representative of a drier environment or at least strong local evaporation. Therefore, it can be proposed that moisture availability was in fact the driving mechanism behind the *ChenoAm*-type-*Podocarpus* relationship.

But what about sea levels?

From the available evidence (e.g. Miller et al. 1995; Jerardino 1997; Baxter & Meadows 1999; Compton 2001) it is indicated that the late Holocene marine transgression along the west and south west coasts of South Africa occurred between c. 4400 and 3800 cal yr BP. This was followed by a regression with present day levels established at around c. 3250 cal yr BP (Meadows et al. 1996; Baxter & Meadows 1999) – which coincides with the onset of the *ChenoAm*-type increase.

In contrast to the initial inferences regarding an increased marine influence at Eilandvlei c. 3200 cal yr BP, it can be suggested that, in addition to greater aridity, the *ChenoAm*-type abundance points to the expansion of salt marshes in and around the lake, as sea- and lake levels decrease.

6.6. The Medieval Climate Anomaly and Little Ice Age in EV1.11

The timing of the global climatic event the Medieval Climate Anomaly (MCA) in South Africa is generally set between c. 1050 and 650 cal yr BP, described as a warmer but also highly variable period (Tyson & Lindesay 1992; Tyson 1997; Tyson & Preston-Whyte 2000; Tyson et al. 2000).

The Little Ice Age (LIA) represents an interval of global cooling, most pronounced in South Africa between c. 650 and 150 cal yr BP (Tyson & Lindesay 1992; Tyson 1997; Tyson & Preston-Whyte 2000; Tyson et al. 2000). The LIA is essentially divided into two cold intervals, punctuated by a ~200 year warmer episode at around c. 450 cal yr BP (Tyson & Lindesay 1992; Tyson 1997; Tyson & Preston-Whyte 2000). The specific timing of this warm event does however vary regionally across South Africa (Tyson & Lindesay 1992).

A closer look at the EV1.1 pollen record for the period c. 1100 to 250 cal yr BP, revealed the trends observed and summarised in Figure 6.10 and Table 6.1.

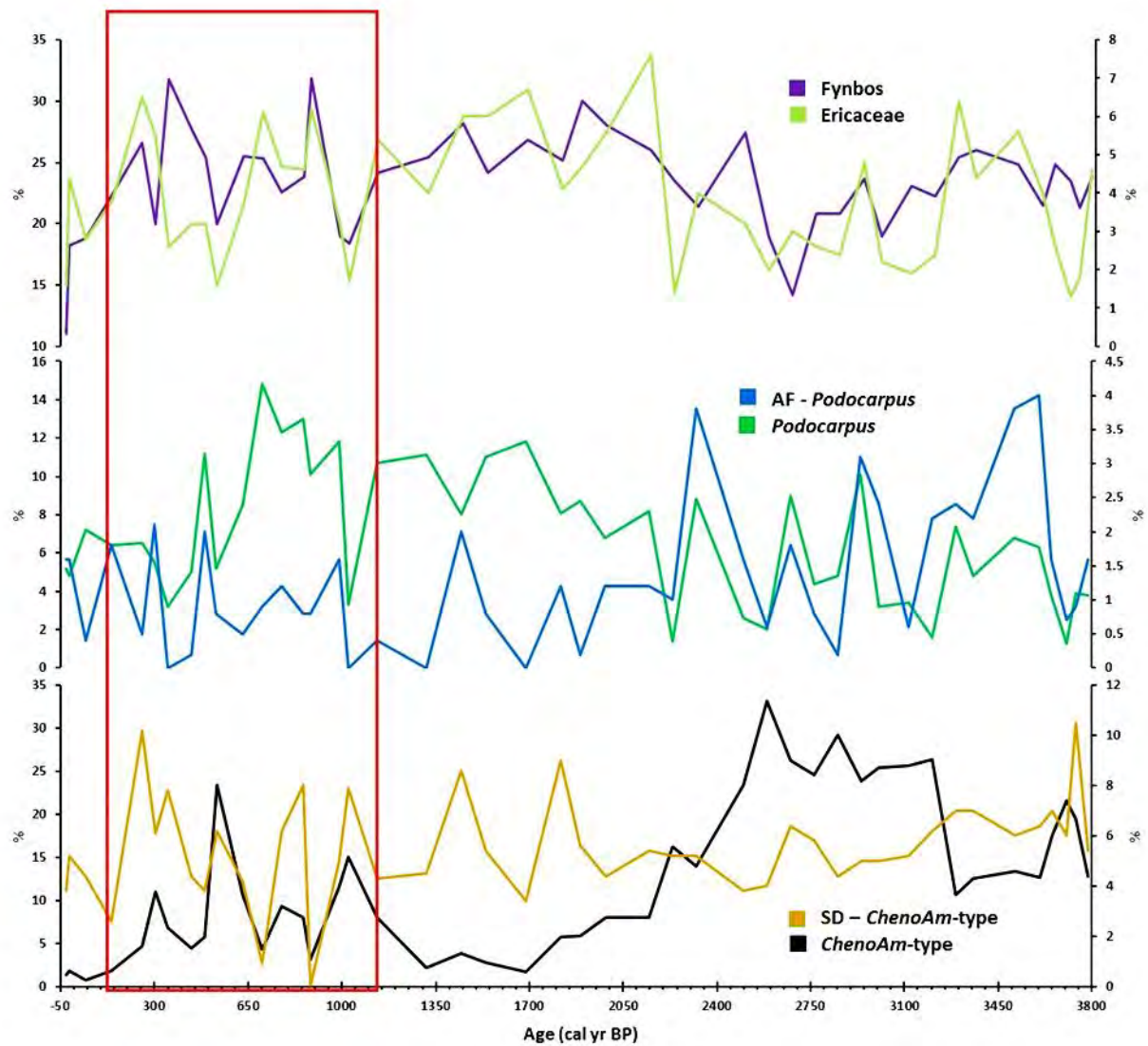


Figure 6.10. Trends in the EV1.11 record for the period c. 1100 to 250 cal yr BP, outlined in red

Table 6.1. A summary of the trends and inferred climatic conditions during the Medieval Climate Anomaly and Little Ice Age from the EV1.11 record

cal yr BP	Trends from EV1.11		Climate	
	<i>Podocarpus</i> and <i>Ericaceae</i> decrease SD and <i>ChenoAm</i> -type peak		drier, most likely warmer	MCA
1100 to 1000				
990	Fynbos increases to maximum <i>Ericaceae</i> increase	Forest peak Aquatics peak	cooler and wetter, rapid increase in moisture availability, winter rainfall seasonality?	First cold phase of LIA
880	SD absent			
850	SD peak			
	SD prompt decline to 700			
700	<i>Podocarpus</i> maximum <i>Anthospermum</i> -type maximum	<i>Ericaceae</i> abundant Fynbos decrease but remains present at >20%	shift to drier, warmer climate	Mid LIA warming
between 700 and 530	<i>Podocarpus</i> marked decline	<i>Ericaceae</i> rapid decrease SD and <i>ChenoAm</i> -type increase		
490	<i>Podocarpus</i> and forest peak	Fynbos increase to near maximum <i>Oxalis</i> present	cool and wet	Second cold phase of LIA
440	<i>Podocarpus</i> and forest decline			
350	<i>Podocarpus</i> only forest element present	<i>Ericaceae</i> increase	trend towards drier, likely warmer environment	Termination of LIA?
300	<i>ChenoAm</i> -type peak			
250	SD peak at near maximum levels	fynbos decline Typha peak		

In section 6.4 the timing of the MCA in the Wilderness Embayment was roughly set for the interval c. 1100 to 700 cal yr BP, largely due to the variability exhibited in the pollen record. As mentioned, Kirsten (2014) suggested the onset of this event at c. 1200 cal yr BP.

A period of increased aridity was identified in the pollen record from around c. 1500 to 1000 cal yr BP, which is further substantiated by the findings from the previous section. Consequently, it appears that the MCA in the Wilderness region was not a rapid event but rather a period of progressively greater aridity, evident in the region from as early as c. 1500 cal yr BP.

The onset of the LIA is not explicitly reflected in the EV1.11 pollen record but according to the patterns observed in the record it was tentatively suggested to be around c. 700 cal yr BP.

Upon further investigation, it can instead be proposed that the LIA started between c. 1000 and 900 cal yr BP as evidenced by the seemingly rapid turnaround from dry and warm conditions to a cooler and wetter environment. These cool and wet conditions prevailed until around c. 700 cal yr BP representing the first cold phase of the LIA.

During the interval c. 700 to around c. 530 cal yr BP, a shift towards a drier and most likely warmer environment is evident in the pollen record, suggesting that this ~200 year event corresponds to the mid-LIA warming phase.

Following this arid episode, a prompt return to wetter and cooler conditions is seen prior to c. 440 cal yr BP. A progressive drying trend is inferred from around c. 440 to 300 cal yr BP while cooler conditions persisted. Indications are thus that the second cool phase of the LIA occurred in the Wilderness region from around c. 530 to 250 cal yr BP.

6.7. Palaeoenvironmental Synthesis

Having explored the EV1.11 pollen record in more detail, it can now be placed in the broader South African context with specific focus on the major trends and events identified in the EV1.11 record.

6.7.1. The Neoglacial and Late Holocene Dune Building

From the onset of the EV1.11 record, c. 3800 cal yr BP, to around c. 3200 cal yr BP the record corresponds to general trends from the WRZ.

A warm and dry environment prevailed in the Wilderness Embayment until c. 3600 cal yr BP with higher temperatures and aridity similarly recorded in the Congo Valley and at Seweweekspoort (Talma & Vogel 1992; Chase et al. 2013). The WRZ likewise experienced warm and dry conditions until around c. 4000 cal yr BP. Some WRZ records do extend this arid period beyond c. 4000 cal yr BP, for example

a drier environment was inferred at Princessvlei until c. 3400 cal yr BP (Kirsten 2014) and at Rietvlei until c. 3300 cal yr BP (Quick et al. in prep).

These drier conditions are most probably a result of the southerly displacement and weakening of the westerlies which causes decreased winter rainfall in the WRZ. Correspondingly, Chase et al. (2013) relate increased aridity at Seweweekspoort to the decreased influence of westerly flow and a resultant decrease in moisture availability.

A cooler climate prevailed throughout the country between c. 4000 and 2000 cal yr BP - this is illustrated best in the Congo Cave temperature record. The record displays a marked decline in temperature from c. 3430 cal yr BP reaching minimum values at c. 3340 cal yr BP and remaining cold until around c. 2600 cal yr BP (Talma & Vogel 1992). While not as pronounced as at Congo Caves, the Cold Air Cave $\delta^{18}\text{O}$ record is punctuated by episodes of significantly decreased $\delta^{18}\text{O}$ levels between c. 3500 and 2500 cal yr BP, which are related to colder temperatures (Holmgren et al. 2003). Lower sea surface temperatures (SST's) were further recorded along both the south- and west coast of South Africa for this period (Cohen et al. 1992; Cohen & Tyson 1995). Cohen & Tyson (1995) suggests that the decreased SST's along the Agulhas Bank was most probably the result of the "intensification of wind-driven coastal upwelling". The timing of this cold interval corresponds with the global cooling event termed the "neo-glacial", characterised by glacial advance in the Northern as well as Southern Hemispheres (Mayewski et al. 2004, and references therein). This event is not explicitly evident in the EV1.11 pollen record although the evidence indicates that cold and possibly wetter conditions prevailed in the region between c. 3600 and 3200 cal yr BP.

During this ~2000 year cool period, the WRZ was generally wetter, due to increased rainfall in the region (e.g. (Meadows & Baxter 2001; Quick et al. 2011; Quick et al. in prep). In accordance with the WRZ, both the Eilandvlei pollen and diatom (Kirsten 2014) records indicate wetter conditions for the century around c. 3650 cal yr BP. A dominant winter rainfall influence and cooler climate was further inferred from the abundance of fynbos, and especially Ericaceae, until c. 3200 cal yr BP. Moisture availability was variable however with forest taxa declining until around c. 3100 cal yr BP. Depleted $\delta^{15}\text{N}$ values from Seweweekspoort, during this period, points to the increasing influence of the westerlies and the possibility of increased humidity (Chase et al. 2013), further substantiating the inference of winter rainfall seasonality in the region. As mentioned previously, Scholtz (1986) inferred summer rainfall seasonality for the South Coast between c. 4400 and 2600 cal yr BP from the increased presence of arboreal taxa. In this regard, Chase and Meadows (2007) suggests that the observed forest expansion in the YRZ after c. 4500 cal yr BP, was instead due to an increased winter rainfall influence as well as lower temperatures.

From around c. 3200 cal yr BP, inferences regarding climate and rainfall seasonality from the EV1.11 pollen record become more tentative. The observed patterns between c. 3200 and 2500 cal yr BP were attributed to greatly enhanced dune activity along the coast rather than a drier environment, as per Martin (1968). This argument still holds but from the findings in section 6.5 it can be posited that increased aridity was in fact experienced in the region during this period. The Eilandvlei diatom record additionally indicates lower lake levels, and thus drier conditions, from c. 2980 cal yr BP (Kirsten 2014).

As this interval corresponds with greater aridity at Seweweekspoort (as seen in section 6.5), it could lead to the conclusion that this drier period is the result of the southerly displacement and reduced influence of the westerlies, as argued by Chase et al. (2013). However, it is postulated that the decrease in moisture availability is instead related to a strengthening of the westerlies, responsible for aeolian deposition, between c. 3000 and 2275 cal yr BP (Carr et al. 2006; Bateman et al. 2011).

The SRZ similarly experienced greater aridity with drier conditions at Lake Eteza between c. 3600 to 2000 cal yr BP (Neumann et al. 2010) and a drier, colder environment at Cold Air Cave from c. 3420 to 2000 cal yr BP (Holmgren et al. 1999; Lee-Thorp et al. 2001).

Hence, it can be proposed that the YRZ tended towards more summer rainfall seasonality, despite the stronger westerlies. The more local environmental disturbances, i.e. dune activity and sea level changes, do however have to be kept in mind, therefore the allusion of summer rainfall seasonality is made very cautiously.

6.7.2. The Medieval Climate Anomaly: c. 1050 to 650 cal yr BP – 900 to 1300 AD

Along the West Coast, evidence suggest largely drier conditions from c. 1050 to 550 cal yr BP (Stager et al. 2012) while at Princessvlei this period tended to be wetter although highly variable (Kirsten 2014). A wetter environment similarly prevailed at Pakhuis Pass and Katbakkies in the Cederberg, from around c. 1000 cal yr BP (Scott & Woodborne 2007b; Chase et al. 2015). The Cold Air Cave records further point to warmer and possibly wetter conditions in the SRZ – brief cool episodes do however occur intermittently (Holmgren et al. 1999; Tyson 1999a, b). The Sibaya diatom record also indicates wetter conditions along the East Coast but Stager et al. (2013) emphasises extremely variable hydroclimatic conditions in the region from c. 950 to 150 cal yr BP. In the Southern Cape interior, the Seweweekspoort $\delta^{15}\text{N}$ record shows a general trend of increasing humidity from c. 1100 to 760 cal yr BP (Chase et al. 2013), with an increase in temperature recorded at Cango Caves c. 1140 to 900 cal yr BP (Talma & Vogel 1992). This points to warm and wet conditions in the region c. 1100 to around 800 cal yr BP.

The drivers behind the climatic trends during the MCA involve the weakening and southerly displacement of the westerly belt - which would have caused the observed drier conditions along the

West Coast (Tyson & Lindesay 1992; Cohen & Tyson 1995). A concurrent intensification in anticyclonic circulation over the South Atlantic and South Indian Oceans would then provide more moisture to the SRZ. (Tyson & Lindesay 1992; Cohen & Tyson 1995). Tyson and Lindesay (1992) further posits that these mechanisms would lead to the dominance of the tropical easterlies along the South Coast, resulting in increased moisture availability in the region.

From the Eilandvlei pollen evidence it is not possible to constrain the MCA in the region to a specific time frame. A definite trend towards greater aridity is however exhibited in the record from c. 1500 to 1000 cal yr BP, from where a prompt shift to a wetter environment occurs, in correspondence with Tyson and Lindesay (1992) as above. This increase in moisture availability is however concomitant with evidence for cooler conditions, i.e. the increased presence of fynbos, especially Ericaceae, which could imply that this interval instead points to the onset of the LIA in the region.

6.7.3. The Little Ice Age: c. 650 to 150 cal yr BP – 1300 to 1850 AD

During this ~500 year period, the SRZ was generally cool and dry as evidenced by the Cold Air Cave records (Tyson 1999a; Holmgren et al. 1999; Lee-Thorp et al. 2001; Holmgren et al. 2003). There are some inconsistencies between the diatom and pollen records from Lake Sibaya, but in general it can be inferred that the East Coast, as with the interior of the SRZ, was colder and drier between c. 650 and 150 cal yr BP (Neumann et al. 2008; Stager et al. 2013).

The WRZ in turn experienced wetter conditions, with Stager et al. (2012) stating that the West Coast was subject to “exceptionally wet conditions during most of the last 7 centuries” (page 884). Baxter and Meadows (1994) similarly note increased moisture availability along the West Coast at this time.

Furthermore, SST's along the West Coast reached minimum values around c. 500 cal yr BP (Cohen et al. 1992) and SST's along the South Coast show a marked decline at c. 600 cal yr BP (Cohen & Tyson 1995).

The Congo Cave temperature record displays colder temperatures during this period with a marked increase in aridity at Seweweekspoort between c. 700 and 570 cal yr BP (Talma & Vogel 1992; Chase et al. 2013), indicating drier and colder conditions, as in the SRZ.

From c. 440 to 330 cal yr BP, a prompt increase in temperature is noted in the Congo Cave record, with a rapid increase in humidity at Seweweekspoort after c. 570 cal yr BP until around c. 430 cal yr BP (Talma & Vogel 1992; Chase et al. 2013). Consequently, this warm and wet interval represents the mid-LIA in the Southern Cape interior.

This phase is also evident in the Cold Air Cave record c. 450 to 275 cal yr BP (Tyson & Lindesay 1992; Tyson 1999b; Holmgren et al. 1999; Tyson et al. 2000). The hiatus in the Princessvlei diatom record,

c. 450 to 320 cal yr BP, is attributed to drier conditions in the area (Kirsten 2014) while this warm, dry phase is further identified at Voëlvlei at around c. 450 cal yr BP (Carr et al. 2006) and at Pearly Beach between c. 450 and 320 cal yr BP (Quick 2013). Stager et al. (2013) additionally notes distinctly wetter conditions along the East Coast between c. 480 and 410 cal yr BP which might be related to the mid LIA event.

Following this warming episode, the lowest temperatures in the Cold Air Cave record is seen at around c. 250 cal yr BP (Tyson & Lindesay 1992; Tyson 1999b; Holmgren et al. 1999; Tyson et al. 2000). Notably drier conditions are evident in the Lake Sibaya diatom record c. 410 to 190 cal yr BP, suggesting that the second phase of the LIA was cold and dry in the SRZ.

The Cango Cave record similarly displays colder temperatures between c. 260 and 170 cal yr BP (Talma & Vogel 1992), although not as pronounced as at Cold Air Cave. The Seweweekspoort $\delta^{15}\text{N}$ record further reaches minimum values at c. 290 cal yr BP (Chase et al. 2013), indicating a cold and wet environment in the Southern Cape during the second cold phase of the LIA.

In the WRZ records, this second phase is less distinct, mostly due to the weaker chronological controls and/ or sampling resolution of the records (e.g. Scott & Woodborne 2007; Meadows et al. 2010; Quick et al. 2011). In addition, the signal associated with anthropogenic disturbances tend to overshadow any climatic or environmental trends which may be present (e.g. Baxter & Meadows 1994; Neumann et al. 2011; Kirsten 2014). Nonetheless, from the Verlorenvlei diatom record, it can be inferred that the WRZ, or at least the West Coast, experienced increased rainfall at this time (Stager et al. 2012).

Drier conditions in the SRZ during the cold phases of the LIA are a result of the increased strength and northward movement of the westerlies, with the concurrent weakening and equatorward movement of anticyclonic circulation, and thus decreased easterly flow (Tyson & Lindesay 1992; Cohen & Tyson 1995; Tyson et al. 2000; Tyson & Preston-Whyte 2000). The increased westerly influence is then responsible for the wetter conditions in the WRZ. Additionally, an expansion of the WRZ towards the interior is suggested during these phases (Tyson & Lindesay 1992). In addition to mechanisms set out above, the coldest episodes during the LIA in South Africa are related to the Maunder and Sporer minima in solar irradiance (Cohen & Tyson 1995; Tyson et al. 2000).

During the mid-LIA warm episode, the opposite would have occurred, i.e. weakening and southerly displacement of the westerlies, with an intensification of the tropical easterlies, resulting in wetter conditions in the SRZ and drier conditions in the WRZ. In accordance with the SRZ, the interior of the Southern Cape was wetter which would point to increased easterly flow in the region due to the contraction of the westerlies and thus the extent of the WRZ.

Climatic trends in the EV1.11 pollen record show general agreement with those of the WRZ, where the first and second phases of the LIA were inferred to be cold and wet along the coast, and the warming phase tended to a drier environment. This can further corroborate the proposed expansion of the WRZ during the cold intervals of the LIA.

According to inferences from the EV1.11 record, conditions along the coast and the interior of the Southern Cape seem to be contrasting. Subsequently it can be proposed that the influence from the easterlies was not as prominent along the coast as in the interior. These disparities could further be attributed to the influence of the Cape Fold Belt- and Outentiqua Mountains separating the two regions, which has a significant effect on the local climate of the Wilderness Embayment.

6.8. A note on the EV1.11 chronology

A possible issue with the chronology of EV1.11 came to light in the previous sections, related to the timing of the MCA and LIA.

Tyson and Lindesay (1992) mention that the environmental expression of these events may not necessarily be contemporaneous, globally or locally, which might explain this inconsistency. There is however a near constant offset of between 200 and 300 years in the timing of the MCA and LIA in EV1.11 in relation to the incidence of these events in South Africa.

The most plausible explanation for this offset would be a marine reservoir effect, which was not taken into account during the calibration of the EV1.11 radiocarbon ages - to date, there is no marine carbon reservoir correction value available for the south coast.

6.9. Conclusion

This chapter explored the EV1.11 record in relation to local, regional and national records. This is essential as Eilandvlei is located within the year round rainfall zone, the climate of which is influenced by climatic mechanisms from both the winter- and summer rainfall zones.

Throughout most of the EV1.11 record, agreement is shown with climatic and environmental trends from the winter rainfall zone, suggesting that variations in the strength and location of the temperate westerlies has a significant effect on the climate of the south coast.

The record further displayed correspondence with local and regional records, i.e. the Eilandvlei diatom record, Groenvlei, Seweweekspoort and Cango Caves, but it was also found that, at times,

conditions along the coast and the interior of the Southern Cape differ significantly – most notably during the Medieval Climate Anomaly and Little Ice Age. This being said, the identification of the MCA and LIA in the EV1.11 record highlighted the possibility of a marine reservoir effect.

7. Conclusion

7.1. Introduction

This study produced a high resolution palaeoenvironmental record of vegetation and climate change during the late Holocene for the Wilderness Embayment; which was achieved through the analysis of the pollen and charcoal data, and results from the Eilandvlei core.

Chapter 7 begins with a summary of the EV1.11 record, focusing on the major events within the record. A review of the aim and objectives for this study are then presented. This is followed by an overview of current and future research in the region.

7.2. Summary of the Findings from the EV1.11 Record

After the in depth analysis of the EV1.11 record, within a local, regional and broader South African context, the following findings warrant emphasis:

- For the period between c. 3800 cal yr BP, the onset of the EV1.11 record, and c. 3200 cal yr BP, conditions at Eilandvlei correspond to those observed within the WRZ, i.e. a warm and dry environment prevailed until around c. 3600 cal yr BP, after which conditions changed to become cooler and most likely wetter.
- Thus, the interval c. 3600 to 3200 cal yr BP most probably represents the expansion of the WRZ – beyond its contemporary extent - indicating that a dominant winter rainfall influence was present along the South Coast.
- A highly variable and dynamic environment characterises the Wilderness Embayment c. 3200 to 2500 cal yr BP, with the area subject to sea level fluctuations, as well as large scale dune activity.
- Dry and windy conditions prevailed in the region during this period; this corresponds with the drier environment in the SRZ, which lead to the very cautious suggestion of a possible shift towards summer rainfall seasonality
- The Medieval Climate Anomaly was recognised as a period when the area experienced progressively greater aridity; this is in contrast to broader trends from the region with increased humidity inferred for the interior of the Southern Cape (Chase et al. 2013). It is further proposed that this increase in moisture availability, both in the interior and along the coast, was the result of dominant easterly flow in the region (Tyson & Lindesay 1992).

- Conditions during the Little Ice Age, as inferred from the EV1.11 record, agree with those of the WRZ, with the first and second phases of the LIA being cold and wet, and the warming phase tending towards greater aridity. Again, this conflicts with the Southern Cape interior.
- In addition to the above disparities between the coast and the interior, a further discrepancy in relation to the timing of the MCA and LIA in the EV1.11 record was identified. Further investigation into the possibility of a marine reservoir effect is needed
- The arrival of European settlers in the region is a further prominent feature in the EV1.11 record marked by the appearance of *Pinus* pollen, c. 1810 AD, and a significant increase in the sedimentation rate.

7.3. Review of the Aim and Objectives

The central aim of this study was to reconstruct a late Holocene record of environmental and climate change for Eilandvlei and its associated catchment, by means of pollen- and microscopic charcoal analysis. Secondary to this, the study was aimed at the investigation of the relative importance of various environmental drivers that may have affected the late Holocene environment in this area; with specific focus on changes in climate, sea level fluctuations and the anthropogenic impact. These aims were reached by achieving the objectives set out in Chapter 1.

Objectives 1 and 2: These objectives were achieved by applying the methodology outlined in Chapter 4, that is, the retrieval of the core, the establishment of a high resolution chronology through radiocarbon dating, and the extraction and identification of pollen and charcoal from these sediments. As presented in Chapter 5, this resulted in the ~3800 year pollen and charcoal records for Eilandvlei.

Objectives 3, 4 and 5: Chapter 6 addresses these objectives, resulting in:

- A detailed ~3800 year palaeoenvironmental reconstruction for Eilandvlei,
- The identification of significant events within the record as well as exploring the climatic mechanisms and environmental drivers behind these
- A synthesis, where the Eilandvlei record is placed within the broader context of the three major rainfall zones of South Africa

7.4. Current and Future Research

The Wilderness Embayment still has much to offer in terms of palaeoenvironmental evidence. Currently, a large multi-proxy research project is underway, focusing on the late Quaternary in the Wilderness Lakes complex. The findings from these studies will undoubtedly contribute greatly to the understanding of climate and environmental dynamics in the region.

With regard to the EV1.11 record, the possibility of a marine reservoir effect does have to be investigated further.

7.5. Conclusion

This study goes a long way in contributing to the current state of knowledge for the Wilderness Embayment, as well as the year round rainfall zone as a whole. It provides a record of vegetation and climate change during the Late Holocene, providing insight into vegetation responses during a highly variable, highly dynamic era.

Finally, in answering the question ‘How can we use the past to predict the future?’, the following quote is deemed appropriate:

“So it was in historic times, still is at present, and will be in the future”

(Tyson & Preston-Whyte 2000, page 305)

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Appendix A

The following steps are needed to create suitable microscopic slides for pollen analysis:

(From Moore et al. 1991; Baxter 1996; Duncan 2006; Quick 2013)

A. Removal of Humic Acids

1. Add 10ml of 10% Sodium Hydroxide (NaOH) to sediment sample and place in water bath for 10 minutes
2. Remove samples and sieve through 160 µm sieve using distilled water
3. Place resultant wash into labelled polypropylene tubes, centrifuge, decant and recombine into single sample
4. Wash in distilled water, centrifuge and decant. Repeat.
5. If supernatant is still darkly coloured, rewash in 10% NaOH and repeat step 4, until supernatant is clear

B. Removal of Clastic Material (Cold Hydrofluoric Acid treatment)

1. Add 20ml of 10% Hydrochloric acid (HCl)
2. Place in water bath for 20 minutes, stir, centrifuge and decant
3. Working in the fume cupboard, add 15ml of 40% Hydrofluoric acid (HF) to each sample, transferring the HF from a container to the sample using a polypropylene syringe to reduce incidence of splashing/spilling
4. Leave samples for 24-48 hours, with the length of the treatment depending on the grittiness of the sample. Stir at regular intervals
5. Seal tubes tightly, centrifuge and decant the supernatant into an appropriate waste container using a polypropylene syringe
6. Add 10ml of 10% HCl, stir, centrifuge and decant
7. Wash in distilled water, centrifuge and decant supernatant
8. Neutralise any spills and ensure that any equipment/surfaces that may have come into contact with HF are thoroughly rinsed, to prevent accidental contact

C. Removal of Cellulose and Organic Matter (Acetolysis Digestion)

1. Working in the fume cupboard, make up the acetolysis mixture
 - 9 parts Acetic Anhydride to 1 part Sulphuric Acid

2. Dehydrate sample by adding 10ml of glacial acetic acid, mix, centrifuge and decant
3. Add 10ml of the acetolysis mixture and place in a water bath for 3 minutes
4. Stop the reaction by adding 5ml of glacial acetic acid. Centrifuge and decant
5. Add 10ml of glacial acetic acid, centrifuge and decant
6. Wash in a mixture of 9ml distilled water and 1ml 10% NaOH to neutralise the sample, centrifuge and decant
7. Wash, twice, in distilled water, adding 1 drop of aqueous safranin stain to the second wash. Centrifuge and decant
8. Add 5ml Tert-butanol alcohol (TBA), stir, centrifuge and decant to just above the pellet
9. Transfer the sample into a labelled storage vial using a micropipette, washing the pipette tip and vial with TBA to ensure thorough transfer of sample material
10. Add equal amount of glycerol to TBA-sample suspension and mix thoroughly to suspend pollen material in glycerol

D. Mounting

1. Place a drop of glycerol onto a warm, labelled sterile glass microscope slide
2. Using the micropipette, extract an appropriate quantity of suspended glycerol-TBA-pollen mixture and release onto the warmed glycerol on the slide
3. Allow TBA content to evaporate
4. Lower the cover slip carefully onto the glycerol-pollen suspension to avoid air bubbles forming, and allow for glycerol to spread evenly to the edges of the slide
5. Using a larger pipette, seal the slide by applying a thin border of molten glycerine jelly (impregnated with phenol) in a steady stream around the edges of the slide
6. Take slide off heat and seal with cover slip

Appendix B: Absolute pollen, spore and micro-charcoal counts

Sample ID	EV1.11-1	EV1.11-3	EV1.11-5	EV1.11-7	EV1.11-9	EV1.11-9.1	EV1.11-10	EV1.11-11	EV1.11-11.1	EV1.11-12.1	EV1.11-13.1	EV1.11-14.1	EV1.11-16	EV1.11-18
Depth (cm)	12	16.5	22	26	30.5	32.5	34.5	38	40	42.5	48	52	56	60.5
Age (cal yr BP)	-28	-16.5	42	141	252.5	302	351.5	438	488	532	631	703	774	855
Artemisia	16	4	3	3	6	5	13	6	5	6	1	1	2	12
Pentzia-type	4	3	6	2	4	4	8	6	7	8	4	0	10	18
Proteaceae	4	2	1	2	0	1	4	2	3	1	0	0	3	2
Ericaceae	8	22	14	19	14	8	13	16	16	8	7	7	23	23
Cliffortia	3	7	5	8	0	0	11	11	4	3	2	1	4	2
Bruniaceae	8	6	0	1	0	0	2	2	6	2	3	0	1	2
Anthospermum-type	7	17	23	22	3	2	11	16	20	15	8	8	15	16
Apiaceae	0	0	0	0	0	0	0	0	2	0	0	0	0	0
Fabaceae	1	0	0	0	0	0	0	0	0	0	0	0	0	0
Rhamnaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Rutaceae	2	6	0	2	0	0	4	2	6	5	1	0	1	0
Santalaceae	0	0	0	1	0	0	0	0	1	3	0	0	0	0
Scroph-type	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Passerina	7	7	8	11	8	2	13	10	18	15	3	1	11	10
Cunoniaceae	1	3	0	0	1	0	0	0	1	1	0	0	0	0
Canthium	3	1	0	1	0	0	0	0	0	0	0	0	0	2
Gnidia	0	0	0	0	0	1	0	0	2	2	0	0	0	1
Euclea	8	7	1	2	5	0	6	13	5	11	1	1	3	6
Grewia	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Ilex	1	1	1	5	0	0	0	0	7	2	1	1	2	4

Sample ID	EV1.11-1	EV1.11-3	EV1.11-5	EV1.11-7	EV1.11-9	EV1.11-9.1	EV1.11-10	EV1.11-11	EV1.11-11.1	EV1.11-12.1	EV1.11-13.1	EV1.11-14.1	EV1.11-16	EV1.11-18
Depth (cm)	12	16.5	22	26	30.5	32.5	34.5	38	40	42.5	48	52	56	60.5
Age (cal yr BP)	-28	-16.5	42	141	252.5	302	351.5	438	488	532	631	703	774	855
Myrica	3	1	0	0	0	0	0	0	0	0	0	0	0	2
Myrsine	2	0	0	0	0	1	0	0	0	0	0	0	2	0
Myrtaceae	3	6	0	1	0	0	0	0	0	0	0	0	0	0
Olea	7	11	8	16	4	7	10	15	18	8	9	3	15	17
Cyperaceae	84	57	47	63	23	14	74	109	55	73	16	12	53	45
Restionaceae	13	9	17	28	13	6	44	66	32	39	17	7	34	49
Poaceae	38	51	49	70	44	27	66	93	76	61	33	22	96	84
Asteraceae HS	15	14	8	8	10	7	11	13	11	7	3	2	10	8
Stoebe-type	1	13	26	17	19	7	52	12	19	11	4	5	14	14
Geraniaceae	0	0	0	0	2	0	6	3	0	1	0	0	0	1
Onagraceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Icacinaeae	0	0	0	1	0	1	0	1	0	0	0	0	2	0
Polygonium	1	1	1	0	1	0	0	0	0	0	0	0	0	0
Acacia	5	7	6	1	0	0	0	0	2	0	1	0	0	0
Celastraceae	3	5	3	6	1	0	7	2	5	1	4	1	6	6
Celtis	1	0	0	0	0	0	0	0	0	0	0	0	0	0
Diospyros	4	5	4	6	0	0	3	0	5	4	1	1	1	2
Euphorbia	4	15	10	7	11	4	18	3	2	15	1	1	11	12
Clutia	3	4	0	2	0	0	0	0	0	0	0	0	0	0
ChenAm-type	7	9	4	9	10	16	34	22	29	117	20	5	45	40

Sample ID	EV1.11-1	EV1.11-3	EV1.11-5	EV1.11-7	EV1.11-9	EV1.11-9.1	EV1.11-10	EV1.11-11	EV1.11-11.1	EV1.11-12.1	EV1.11-13.1	EV1.11-14.1	EV1.11-16	EV1.11-18
Depth (cm)	12	16.5	22	26	30.5	32.5	34.5	38	40	42.5	48	52	56	60.5
Age (cal yr BP)	-28	-16.5	42	141	252.5	302	351.5	438	488	532	631	703	774	855
Crassula	1	4	1	2	3	0	6	6	4	3	0	0	5	8
Gunneraceae	0	0	0	1	0	0	0	0	0	0	0	0	0	0
Juncaceae	8	9	4	9	0	0	10	11	10	4	2	0	8	4
Liliaceae	5	4	2	5	2	0	9	6	7	6	3	3	7	4
Iridaceae	2	2	0	2	0	3	5	2	3	1	3	0	4	1
Urticaceae	0	3	2	0	0	0	0	0	3	0	0	0	1	0
Verbenaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Haloraginataceae	0	0	0	0	0	0	0	0	1	0	0	0	1	0
Typha	3	2	3	1	0	3	0	0	1	0	0	1	4	3
Zygophyllaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Lauraceae	1	0	1	1	0	0	0	0	2	1	0	0	0	0
Loranthaceae	2	0	0	0	0	0	0	0	0	0	0	0	0	0
Plantago	1	4	4	4	0	0	0	0	3	0	0	0	0	1
Acanthaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Pinaceae	123	76	168	71	0	0	0	0	0	0	0	0	0	0
Podocarpus	26	24	36	32	14	8	16	25	56	26	16	17	60	65
Rhus	6	5	6	10	2	4	4	5	11	9	4	5	3	5
Aizoaceae	10	4	5	2	2	1	1	4	6	4	3	0	4	1
Euphorbiaceae undif	0	0	0	0	3	3	4	3	1	2	0	0	4	4
Brassicaceae	5	9	1	2	2	2	8	3	5	4	1	0	5	1
Campanulaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Caryophyllaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	2
Moraceae	0	0	0	0	0	1	0	0	0	0	0	0	0	0

Sample ID	EV1.11-1	EV1.11-3	EV1.11-5	EV1.11-7	EV1.11-9	EV1.11-9.1	EV1.11-10	EV1.11-11	EV1.11-11.1	EV1.11-12.1	EV1.11-13.1	EV1.11-14.1	EV1.11-16	EV1.11-18
Depth (cm)	12	16.5	22	26	30.5	32.5	34.5	38	40	42.5	48	52	56	60.5
Age (cal yr BP)	-28	-16.5	42	141	252.5	302	351.5	438	488	532	631	703	774	855
Polygala	0	0	0	2	0	0	0	0	1	0	0	0	2	0
Oxalis	0	0	2	2	0	0	3	1	2	4	0	0	2	2
Unidentifiable	19	28	8	18	2	3	9	4	12	4	1	1	4	8
Broken	16	26	10	19	3	3	11	7	13	10	6	1	8	11
Unknown	5	6	2	3	2	2	4	0	2	3	9	8	0	2
Total Pollen	500	500	500	500	214	146	500	500	500	500	188	115	486	500
AP	194	151	234	152	27	22	46	61	112	63	37	29	94	107
NAP	306	349	266	348	187	124	454	439	388	437	151	86	392	390
All total	1481	1506	1364	2087	4014	1478	3727	2256	2580	5773	1605	1243	2869	1808
Exotics	143	90	93	53	113	43	275	112	87	322	32	37	75	92
10 - 50 um	795	892	720	1498	3669	1278	2917	1632	1966	4931	1366	1086	2281	1167
50 - 100 um	1	4	6	5	0	0	2	0	4	2	5	3	7	10
Monolete Spore	37	15	41	28	17	10	26	12	20	14	13	2	17	36
Trilete Spore	5	5	4	3	1	1	7	0	3	4	1	0	3	6

Sample ID	EV1.11-19	EV1.11-21	EV1.11-22	EV1.11-25	EV1.11-26	EV1.11-27	EV1.11-29	EV1.11-30	EV1.11-32	EV1.11-34	EV1.11-35	EV1.11-36	EV1.11-37	EV1.11-38
Depth (cm)	62	68	70	76	80	82	85.5	88	92	95.5	97.5	100	104	106
Age (cal yr BP)	882	989	1025	1133	1243	1317	1446.5	1539	1688	1817.5	1891.5	1984	2147	2235
Artemisia	2	13	5	6	2	9	14	15	0	9	13	16	12	5
Pentzia-type	0	7	5	3	1	5	22	9	0	6	8	2	0	0
Proteaceae	0	1	0	0	1	3	2	2	1	2	3	5	1	1
Ericaceae	8	16	3	12	2	15	30	30	8	14	23	28	38	4
Cliffortia	0	1	2	2	0	2	7	9	1	7	6	13	6	5
Bruniaceae	0	1	0	2	1	2	8	1	0	3	2	1	2	1
Anthospermum-type	5	6	3	8	3	16	25	8	5	10	32	3	6	18
Apiaceae	0	0	0	0	1	1	0	0	0	0	0	0	0	0
Fabaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Rhamnaceae	0	0	0	0	1	0	1	0	0	0	0	0	0	0
Rutaceae	0	3	4	2	1	3	3	2	0	0	1	7	3	3
Santalaceae	0	0	0	0	0	0	1	0	0	0	0	0	0	1
Scroph-type	0	0	0	0	0	0	0	0	0	0	0	2	0	0
Passerina	6	12	2	4	2	15	19	9	2	8	11	8	17	6
Cunoniaceae	0	0	0	0	0	0	1	1	0	0	0	0	1	0
Canthium	0	0	0	0	0	1	4	0	0	0	0	0	2	0
Gnidia	0	3	0	0	1	2	2	0	0	0	0	1	1	0
Euclea	0	9	4	1	0	5	10	11	2	5	8	6	10	9
Grewia	0	0	0	0	0	0	0	0	0	1	0	0	0	0
Ilex	1	7	0	1	0	0	2	2	0	1	0	1	3	3

Sample ID	EV1.11-19	EV1.11-21	EV1.11-22	EV1.11-25	EV1.11-26	EV1.11-27	EV1.11-29	EV1.11-30	EV1.11-32	EV1.11-34	EV1.11-35	EV1.11-36	EV1.11-37	EV1.11-38
Depth (cm)	62	68	70	76	80	82	85.5	88	92	95.5	97.5	100	104	106
Age (cal yr BP)	882	989	1025	1133	1243	1317	1446.5	1539	1688	1817.5	1891.5	1984	2147	2235
Myrica	0	2	0	0	0	0	0	3	0	0	0	0	0	1
Myrsine	0	0	0	0	0	0	0	0	0	1	0	2	0	0
Myrtaceae	0	0	0	1	0	0	0	1	1	0	0	0	1	0
Olea	5	21	11	9	2	16	35	15	8	17	17	18	25	15
Cyperaceae	12	46	37	16	10	36	42	51	4	52	49	48	60	44
Restionaceae	14	38	13	15	8	22	24	38	9	25	36	53	43	23
Poaceae	28	85	24	45	23	70	39	83	24	36	78	88	93	21
Asteraceae														
HS	1	11	7	5	2	16	25	20	2	9	18	7	9	8
Stoebe-type	7	14	5	9	4	16	20	20	5	18	31	18	14	5
Geraniaceae	0	2	2	1	0	1	0	2	2	9	1	2	1	4
Onagraceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Icacinaeae	0	0	0	0	0	0	7	1	0	1	0	3	0	0
Polygonium	0	0	0	0	0	0	0	0	0	1	1	0	0	0
Acacia	0	0	0	0	0	1	0	0	0	0	0	0	0	0
Celastraceae	1	5	0	2	1	6	14	6	1	3	3	4	3	2
Celtis	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Diospyros	0	3	0	0	1	3	2	5	0	2	3	7	1	2
Euphorbia	0	11	5	3	0	5	10	12	0	11	10	15	18	8
Clutia	0	1	0	0	0	0	0	0	0	0	1	0	0	0
ChenAm-type	4	58	27	18	3	8	19	14	2	20	29	40	40	47

Sample ID	EV1.11-19	EV1.11-21	EV1.11-22	EV1.11-25	EV1.11-26	EV1.11-27	EV1.11-29	EV1.11-30	EV1.11-32	EV1.11-34	EV1.11-35	EV1.11-36	EV1.11-37	EV1.11-38
Depth (cm)	62	68	70	76	80	82	85.5	88	92	95.5	97.5	100	104	106
Age (cal yr BP)	882	989	1025	1133	1243	1317	1446.5	1539	1688	1817.5	1891.5	1984	2147	2235
Crassula	0	3	1	2	0	2	11	1	0	1	5	1	5	3
Gunneraceae	0	0	0	0	0	0	0	1	0	0	0	0	0	0
Juncaceae	1	6	0	2	1	5	3	5	2	1	6	0	2	1
Liliaceae	5	5	3	1	0	4	5	5	3	4	5	8	1	8
Iridaceae	1	3	1	0	0	1	2	2	1	0	3	4	0	2
Urticaceae	0	0	0	1	0	0	0	0	0	0	1	0	0	0
Verbenaceae	0	0	0	0	0	0	0	0	0	0	0	0	1	0
Haloraginatae	0	0	0	0	0	0	1	0	0	0	0	0	1	0
Typha	1	8	0	0	2	3	9	11	5	0	0	8	4	0
Zygophyllaceae	0	0	0	0	0	1	0	0	0	0	0	0	0	0
Lauraceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Loranthaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Plantago	0	0	0	1	0	1	0	0	0	0	0	0	0	0
Acanthaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Pinaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Podocarpus	13	59	6	24	6	41	40	55	14	28	43	34	41	4
Rhus	6	8	0	8	1	6	19	4	2	14	6	13	8	12
Aizoaceae	0	2	1	1	1	3	0	3	2	4	4	2	3	0
Euphorbiaceae undif	0	7	0	0	0	0	0	9	0	4	2	6	9	1
Brassicaceae	0	4	4	3	0	5	8	6	4	5	5	4	2	6
Campanulaceae	0	0	0	0	0	0	1	0	0	0	0	0	0	0
Caryophyllaceae	0	0	0	0	0	1	0	0	0	0	0	0	0	0
Moraceae	0	0	0	0	0	0	0	0	0	0	0	0	2	0

Sample ID	EV1.11-19	EV1.11-21	EV1.11-22	EV1.11-25	EV1.11-26	EV1.11-27	EV1.11-29	EV1.11-30	EV1.11-32	EV1.11-34	EV1.11-35	EV1.11-36	EV1.11-37	EV1.11-38
Depth (cm)	62	68	70	76	80	82	85.5	88	92	95.5	97.5	100	104	106
Age (cal yr BP)	882	989	1025	1133	1243	1317	1446.5	1539	1688	1817.5	1891.5	1984	2147	2235
Polygala	1	2	0	0	0	0	0	2	0	0	1	1	1	0
Oxalis	0	0	0	1	0	0	0	2	0	1	2	1	0	0
Unidentifiable	0	9	0	4	1	10	4	7	2	7	9	6	6	6
Broken	3	6	3	4	0	4	8	11	2	4	7	8	2	4
Unknown	4	2	2	7	0	5	1	6	5	1	10	6	2	5
Total Pollen	129	500	180	224	82	371	500	500	119	345	493	500	500	288
AP	26	114	21	46	11	78	130	104	28	73	80	88	95	48
NAP	103	383	159	178	71	293	370	396	91	272	413	412	406	240
All total	2847	2699	2069	3044	1784	2942	2194	3293	4449	4311	12984	6140	2117	6566
Exotics	69	116	132	77	26	118	84	139	101	152	357	143	131	265
10 - 50 um	2629	2041	1740	2723	1670	2420	1582	2615	4217	3801	12077	5462	1416	5981
50 - 100 um	7	4	7	2	0	3	3	3	3	2	13	13	13	12
Monolete Spore	11	34	7	16	5	24	14	33	8	8	38	19	43	12
Trilete Spore	2	7	3	2	1	6	10	3	1	3	6	3	13	8

Sample ID	EV1.11-39	EV1.11-40	EV1.11-40.1	EV1.11-41	EV1.11-42	EV1.11-43	EV1.11-44	EV1.11-45	EV1.11-46	EV1.11-47	EV1.11-48	EV1.11-49	EV1.11-50	EV1.11-51
Depth (cm)	108	112	114	116	118	120	122	123.5	126	128	130	131.5	136	138
Age (cal yr BP)	2323	2499	2587	2675	2763	2851	2939	3005	3116	3204	3292	3358	3513	3559
Artemisia	7	6	6	7	9	6	6	8	3	7	11	8	21	4
Pentzia-type	5	6	1	10	4	3	3	3	1	0	5	8	3	1
Proteaceae	4	3	0	1	2	1	3	3	0	2	6	5	3	1
Ericaceae	20	16	7	15	13	12	17	11	8	12	32	22	28	8
Cliffortia	1	7	2	2	3	4	5	2	1	4	3	5	11	0
Bruniaceae	1	3	2	0	2	0	4	2	3	5	7	7	0	1
Anthospermum-type	14	13	6	7	8	9	6	3	8	12	20	6	6	1
Apiaceae	0	0	1	0	0	0	0	0	0	1	0	0	0	0
Fabaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Rhamnaceae	0	0	0	0	0	0	0	1	0	0	0	0	0	0
Rutaceae	2	4	1	1	3	2	1	11	3	2	2	4	5	0
Santalaceae	0	0	2	0	0	1	0	1	2	0	0	0	2	0
Scroph-type	0	1	0	0	1	0	0	0	0	0	0	0	1	0
Passerina	7	16	8	5	15	11	7	7	11	14	9	18	13	1
Cunoniaceae	0	0	0	0	0	0	1	0	0	0	0	0	0	0
Canthium	0	1	0	0	1	0	0	0	0	0	0	0	2	0
Gnidia	0	1	2	1	3	4	0	3	2	0	1	0	2	0
Euclea	12	8	5	7	10	9	5	8	8	11	17	7	10	2
Grewia	0	2	0	0	0	0	0	2	0	6	0	1	4	0
Ilex	5	3	1	0	2	0	4	2	1	2	6	4	8	0

Sample ID	EV1.11-39	EV1.11-40	EV1.11-40.1	EV1.11-41	EV1.11-42	EV1.11-43	EV1.11-44	EV1.11-45	EV1.11-46	EV1.11-47	EV1.11-48	EV1.11-49	EV1.11-50	EV1.11-51
Depth (cm)	108	112	114	116	118	120	122	123.5	126	128	130	131.5	136	138
Age (cal yr BP)	2323	2499	2587	2675	2763	2851	2939	3005	3116	3204	3292	3358	3513	3559
Myrica	0	0	1	0	1	0	0	0	1	1	0	1	1	0
Myrsine	3	3	0	0	0	0	0	0	1	3	0	2	0	0
Myrtaceae	0	0	0	0	0	0	0	1	0	2	0	0	0	0
Olea	11	8	5	17	19	14	19	15	10	15	23	18	8	4
Cyperaceae	35	49	66	36	50	59	24	69	52	59	48	56	39	5
Restionaceae	30	51	31	21	35	43	32	31	40	40	28	37	41	8
Poaceae	85	41	29	69	60	46	32	49	37	30	65	77	69	8
Asteraceae														
HS	7	4	1	3	11	7	2	7	3	8	10	7	7	1
Stoebe-type	24	16	5	18	19	17	7	20	16	17	16	23	17	5
Geraniaceae	3	3	6	5	0	2	0	8	4	3	6	1	2	0
Onagraceae	0	0	0	0	0	0	0	0	0	0	2	0	0	0
Icacinaeae	3	0	1	1	0	1	1	3	1	0	4	4	4	1
Polygonium	0	0	0	0	0	0	0	0	0	1	0	0	0	0
Acacia	2	0	0	0	0	0	0	0	0	0	0	0	0	0
Celastraceae	5	3	2	7	3	0	4	1	3	0	4	3	3	0
Celtis	0	3	0	0	0	0	1	0	0	1	0	0	0	0
Diospyros	3	2	2	1	2	0	2	2	2	1	1	6	3	0
Euphorbia	13	8	5	13	14	10	9	10	8	16	17	14	17	4
Clutia	1	0	0	0	0	0	0	0	0	0	0	0	0	0
ChenAm-type	70	117	119	131	123	146	85	127	107	132	53	63	67	19

Sample ID	EV1.11-39	EV1.11-40	EV1.11-40.1	EV1.11-41	EV1.11-42	EV1.11-43	EV1.11-44	EV1.11-45	EV1.11-46	EV1.11-47	EV1.11-48	EV1.11-49	EV1.11-50	EV1.11-51
Depth (cm)	108	112	114	116	118	120	122	123.5	126	128	130	131.5	136	138
Age (cal yr BP)	2323	2499	2587	2675	2763	2851	2939	3005	3116	3204	3292	3358	3513	3559
Crassula	5	2	2	2	8	0	4	2	3	5	4	8	5	1
Gunneraceae	0	0	0	0	0	0	0	0	0	0	0	0	2	0
Juncaceae	6	0	0	3	2	1	2	5	3	3	0	2	2	0
Liliaceae	4	13	7	7	5	10	2	6	8	8	5	5	2	1
Iridaceae	4	8	5	1	4	5	2	4	7	3	4	3	0	0
Urticaceae	0	1	0	0	0	0	0	0	0	0	0	0	1	0
Verbenaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Haloraginataceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Typha	6	6	1	5	4	3	0	0	1	3	7	0	7	0
Zygophyllaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Lauraceae	0	0	0	1	2	0	0	0	0	0	2	0	0	0
Loranthaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Plantago	0	0	0	1	1	0	1	0	0	0	0	0	0	0
Acanthaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Pinaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Podocarpus	44	13	7	45	22	24	36	16	14	8	37	24	34	9
Rhus	6	14	5	10	10	8	5	10	2	17	14	16	19	4
Aizoaceae	0	0	0	2	3	7	2	2	6	7	3	4	3	0
Euphorbiaceae undif	5	3	0	6	8	3	4	5	2	5	5	6	1	0
Brassicaceae	4	7	3	5	5	9	0	9	9	5	6	6	7	1
Campanulaceae	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Caryophyllaceae	5	1	0	3	0	0	0	0	0	0	0	0	0	0
Moraceae	7	0	0	7	0	0	5	5	0	0	0	0	3	0

Sample ID	EV1.11-39	EV1.11-40	EV1.11-40.1	EV1.11-41	EV1.11-42	EV1.11-43	EV1.11-44	EV1.11-45	EV1.11-46	EV1.11-47	EV1.11-48	EV1.11-49	EV1.11-50	EV1.11-51
Depth (cm)	108	112	114	116	118	120	122	123.5	126	128	130	131.5	136	138
Age (cal yr BP)	2323	2499	2587	2675	2763	2851	2939	3005	3116	3204	3292	3358	3513	3559
Polygala	0	2	1	1	1	0	0	0	1	2	2	0	0	0
Oxalis	0	0	0	0	0	1	0	0	2	0	0	0	0	0
Unidentifiable	11	17	2	6	5	9	5	10	7	8	9	8	7	1
Broken	10	7	4	7	3	7	5	11	8	10	4	7	6	1
Unknown	10	8	4	10	4	6	2	5	7	9	2	4	4	1
Total Pollen	500	500	358	500	500	500	355	500	416	500	500	500	500	93
AP	101	59	29	96	71	56	83	65	43	67	109	86	97	20
NAP	399	441	329	404	429	444	272	435	373	433	387	414	403	73
All total	10554	4812	4062	9670	3050	4548	8469	3274	4965	4689	10160	3459	4382	4639
Exotics	201	377	254	295	121	403	88	226	296	324	154	155	117	74
10 - 50 um	9796	3904	4185	8827	2409	3625	7994	2530	4223	3838	9438	2777	3738	4451
50 - 100 um	9	12	14	10	0	6	11	5	13	12	26	0	3	11
Monolete Spore	42	8	10	33	16	11	17	13	10	10	35	22	18	8
Trilete Spore	6	11	4	5	4	3	4	0	7	5	11	5	6	2

Sample ID	EV1.11-52	EV1.11-53	EV1.11-54	EV1.11-55	EV1.11-56	Sample ID	EV1.11-52	EV1.11-53	EV1.11-54	EV1.11-55	EV1.11-56
Depth (cm)	140	142	144.5	146	148	Depth (cm)	140	142	144.5	146	148
Age (cal yr BP)	3605	3650	3707	3741	3787	Age (cal yr BP)	3605	3650	3707	3741	3787
Artemisia	1	3	2	0	11	Myrica	1	0	0	1	2
Pentzia-type	0	6	1	5	2	Myrsine	1	0	0	2	0
Proteaceae	0	2	0	4	3	Myrtaceae	0	0	0	0	0
Ericaceae	5	13	2	6	23	Olea	6	13	7	18	14
Cliffortia	0	7	2	0	4	Cyperaceae	7	65	18	40	46
Bruniaceae	1	5	0	1	6	Restionaceae	10	45	16	23	36
Anthospermum-type	1	11	4	4	11	Poaceae	16	43	12	29	83
Apiaceae	0	0	0	1	0	Asteraceae					
Fabaceae	0	0	0	0	0	HS	3	15	2	4	11
Rhamnaceae	0	0	0	0	0	Stoebe-type	7	19	8	19	20
Rutaceae	1	3	0	1	4	Geraniaceae	0	3	2	3	2
Santalaceae	0	0	0	0	0	Onagraceae	0	0	0	0	0
Scroph-type	0	1	0	0	0	Icacinaeae	0	0	0	0	3
Passerina	1	14	4	9	9	Polygonium	0	0	1	0	0
Cunoniaceae	0	1	0	0	0	Acacia	0	0	0	0	0
Canthium	0	0	0	0	0	Celastraceae	2	1	0	0	1
Gnidia	0	1	0	0	1	Celtis	0	2	2	1	0
Euclea	4	12	3	10	15	Diospyros	2	7	2	6	2
Grewia	0	5	0	0	0	Euphorbia	4	16	5	13	17
Ilex	0	2	0	1	3	Clutia	0	0	1	0	2
						ChenAm-type	16	88	33	65	64

Sample ID	EV1.11-52	EV1.11-53	EV1.11-54	EV1.11-55	EV1.11-56	Sample ID	EV1.11-52	EV1.11-53	EV1.11-54	EV1.11-55	EV1.11-56
Depth (cm)	140	142	144.5	146	148	Depth (cm)	140	142	144.5	146	148
Age (cal yr BP)	3605	3650	3707	3741	3787	Age (cal yr BP)	3605	3650	3707	3741	3787
Crassula	2	3	1	4	6	Polygala	0	0	0	0	1
Gunneraceae	1	1	0	0	1	Oxalis	1	1	0	0	0
Juncaceae	0	3	0	3	3	Unidentifiable	2	15	4	6	10
Liliaceae	2	8	4	7	5	Broken	2	10	2	5	8
Iridaceae	1	5	0	4	3	Unknown	3	5	5	4	6
Urticaceae	0	0	0	0	0	Total Pollen	126	500	153	334	500
Verbenaceae	0	0	0	0	0						
Haloragaceae	0	0	0	0	0	AP	33	72	18	58	80
Typha	1	1	1	0	6	NAP	93	428	135	276	420
Zygophyllaceae	0	0	0	0	0	All total	6378	6076	5402	5507	7196
Lauraceae	0	0	0	0	0	Exotics	93	258	226	201	193
Loranthaceae	0	0	0	0	0						
Plantago	0	0	0	0	0	10 - 50 um	6143	5273	5006	4942	6462
Acanthaceae	0	1	0	0	0	50 - 100 um	2	21	10	18	13
Pinaceae	0	0	0	0	0						
Podocarpus	8	19	2	13	19	Monolete Spore	13	20	7	8	21
Rhus	5	10	2	6	21	Trilete Spore	1	4	0	4	7
Aizoaceae	2	7	0	10	0						
Euphorbiaceae											
undif	2	1	1	1	9						
Brassicaceae	0	7	4	5	6						
Campanulaceae	0	0	0	0	0						
Caryophyllaceae	1	0	0	0	1						
Moraceae	4	0	0	0	0						

